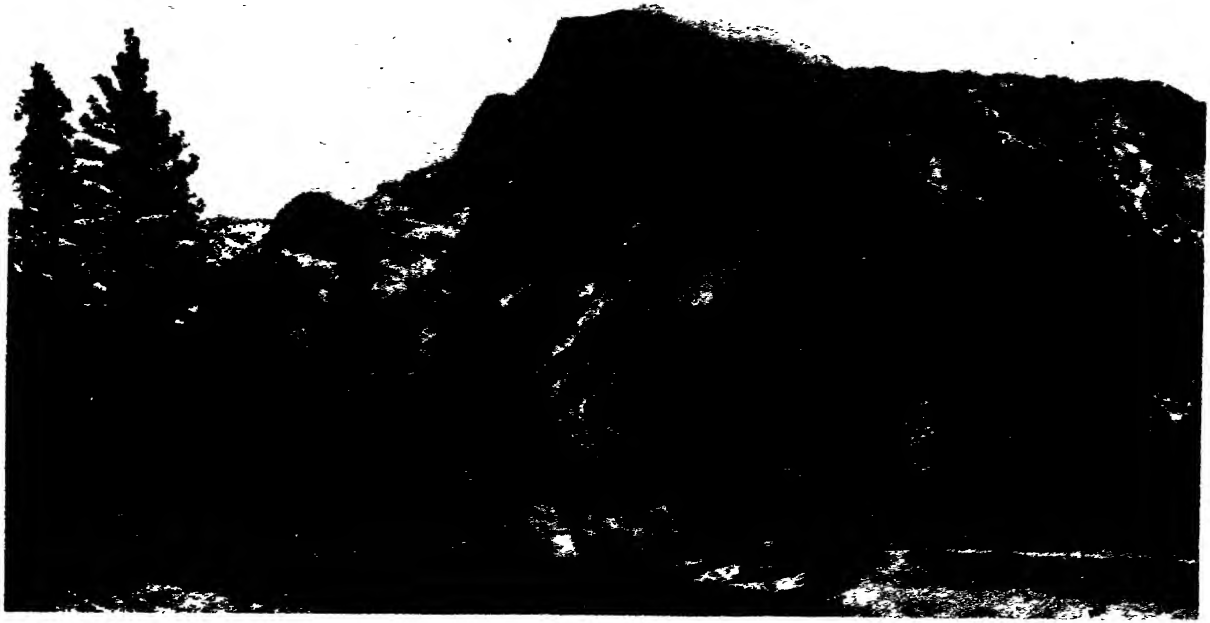
An Introduction to
Physical Geology



HALF DOME IN YOSEMITE NATIONAL PARK, CALIFORNIA.

Rises nearly 4,800 feet above the river. The rock is granodiorite which, in a molten condition, was forced into the earth's crust millions of years ago. By the erosive action of streams the overlying rock material was removed, laying bare the granodiorite. Continued stream erosion, profound erosion by a glacier which almost over-topped the great rock, and some post-Glacial weathering have sculptured the mass into the remarkable form of the present day. (Courtesy of the Southern Pacific Lines.)

An Introduction to
Physical Geology

By

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CHAPTER I

INTRODUCTION

GEOLOGY—A BROAD SCIENCE

ALTHOUGH the significance of the physical world to the existence of mankind is never questioned, an appreciation and understanding of it are pretty much lacking by the public in general. Too commonplace are the misimpressions and lack of knowledge of such matters as history of the earth, the conditions of the earth's interior, the materials of the earth, the nature and development of the landscape, the forces ever at work to bring about constant change, and even the atmosphere which man must breathe. It is the science of *geology* which reveals so much knowledge of this physical world. In simplest terms geology is defined as the study of the earth. The word is derived from the Greek *ge* meaning "earth," and *logia,* meaning "discourse." By its very nature, the science of geology enlists the use of certain phases of many other sciences, especially physics, chemistry, botany, zoology, and astronomy, thus making it one of the broadest and most comprehensive of scientific subjects. Broadly considered, it may be divided into two general subdivisions, (a) *physical geology* and (b) *historical geology*. *Physical geology* is concerned with the materials and structure of the earth, the configuration of the earth's surface, and forces by which the earth has been for many millions of years, and is being, modified both internally and externally. It is, therefore, concerned simply with the physical aspects of the earth and its behavior. *Historical geology,* on the other hand, deals with the succession of historical events of the earth and its inhabitants as revealed through the record in the rocks of the earth's crust. It is concerned with the succession of both physical events and life events in the far distant past, as deduced through the study of the complex assemblage of rock layers of the earth's crust which constitute the pages of nature in the record of the earth's past. This text deals with physical geology only.

The Earth as a Planet

Since the earth is the subject of geology, it is important that one have clearly in mind certain well-known facts regarding it as a planet. The earth is a member of the so-called solar system of which the sun, whose diameter is about 866,000 miles, is the center. Nine planets, including the earth, revolve in the same direction but along different orbits in approximately the same plane around the sun. The earth's form is that of an oblate spheroid. Its average diameter, excluding its atmosphere, is nearly 8000 miles. From pole to pole it is 27 miles shorter than across the equatorial plane. The earth rotates on its axis once in 24 hours, thus defining a day. Its mean distance from the sun is nearly 93,000,000 miles, and it revolves around the sun once in 365¼ days. Certain planets are much larger and farther away from the sun than the earth, whereas some others are smaller, two of them being nearer the sun.

One satellite, called the moon, revolves around the earth once in about 28 days at a distance of 240,000 miles. Although the moon is much smaller than the earth, it nevertheless has some indirect geological influence upon the earth since it is the principal cause of ocean tides. Other planets are also attended by similar satellites in varying numbers. The geological influence of the sun upon the earth is far greater than that of the moon because the sun is the chief source of the earth's light, heat, and energy, which have made largely or wholly possible many geologic activities.

Major Divisions of the Earth

The three great divisions of the earth are (a) the lithosphere, (b) the hydrosphere, and (c) the atmosphere. They are complexly interrelated and are not independent of one another.

Lithosphere. This term (from the Greek *lithos,* meaning "stone") in its broadest concept refers to the main, rock body of the earth which is largely in the solid state, and which extends from the soil on the surface to possibly the center.

There is now a tendency to limit the lithosphere to the outer, stronger, rocky (crystalline) shell beneath which there is a weaker, earth-circling shell, called the asthenosphere, which may or may not be crystalline, and which in turn envelops a large central mass or core of unknown material with still different properties.

Although only a thin, outermost shell of rock, called the *crust,* has come under the direct observation of man, the earth's interior is known, through deduction from the behavior of earthquake waves, to be comprised of a series of concentric shells of different rock compositions and physical properties surrounding a heavy, central core of over 4000 miles in diameter which may be composed of nickel and iron. The crustal part of the earth is of supreme importance to the geologist not only because of its minerals and rocks of economic value but because it possesses the record of the wonderful events of earth history.

Hydrosphere. The partial envelope comprising all the waters on and near the earth's surface is called the hydrosphere. Most of the waters by far are in the oceans; but streams, lakes, and underground waters are also important. Water is one of the greatest of all the geological agencies which, for countless ages, have been modifying the earth. Its greatest function is the wearing down, through stream erosion, of the higher portions of the land, and the transportation and deposition of the resulting sediment in the various lower portions (basins) of the earth's surface.

Atmosphere. The gaseous envelope of the earth is called the atmosphere (air). To some extent it penetrates the outer portion of the earth's crust through openings in the rocks, and to some extent it is dissolved in the waters. The chief constituents of the lower air are nitrogen (78%), oxygen (21%), water vapor, argon, and carbon dioxide, most of which effect important changes in the rocks and minerals of the earth. Although rare atmospheric gases extend to heights of hundreds of miles above sea level, one half of the entire mass of air rests below an elevation of 3½ miles. The atmosphere exerts a pressure of 14.7 pounds per square inch on the surface of the lithosphere at sea level. Movements of the atmosphere (winds) cause important modifications of the lands, especially in arid regions. Probably the greatest geological function of the atmosphere is an indirect one—its making possible precipitation (rainfall and snowfall), which in turn makes possible the work of running water and glaciers.

The Scope and Significance of Geology

Important conclusions. Unless he has devoted some study to the matter, the average person is very likely to regard the great variety of physical features and life of the earth as practically unchangeable, and

to think that they were essentially the same in the beginning of the earth's history as they are now. But the study of geology has firmly established the great fact that the face of the earth and the life upon it as we see it today represent merely a single phase of a tremendously long history which has involved many profound and far-reaching changes.

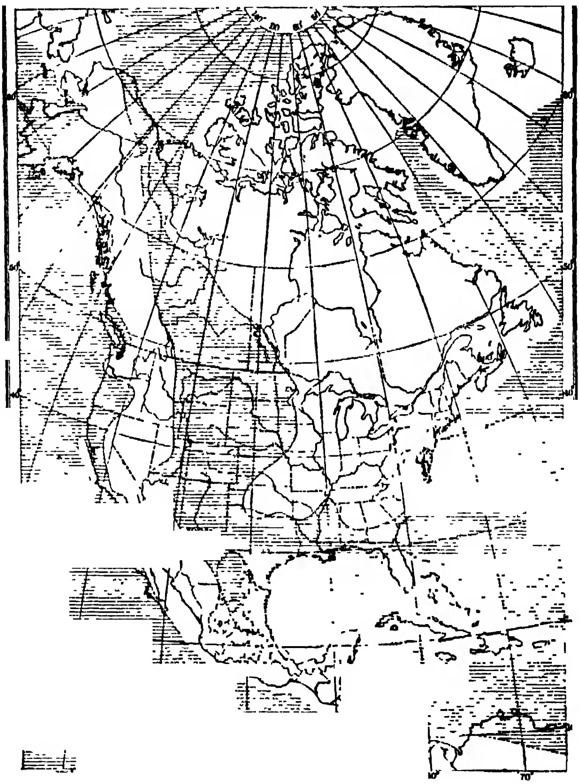


FIG. 1. Map of North America showing the spread of the sea over the land during part of Cretaceous time, millions of years ago. (After C. Schuchert.)

The following concise statements of some of the more definite and important conclusions regarding earth changes may serve to give a fair conception of the general scope and significance of geology. For untold millions of years rocks at and near the surface of the earth have been

short as compared to that of known geological time. The one is to be measured by thousands of years, and the other by hundreds of millions of years. To the geologist a lapse of hundreds of thousands of years is a "short" time. "The flowing landscapes of geologic time may be likened to a kinetoscopic panorama. The scenes transform from age to age; seas and plains and mountains of different types follow and replace

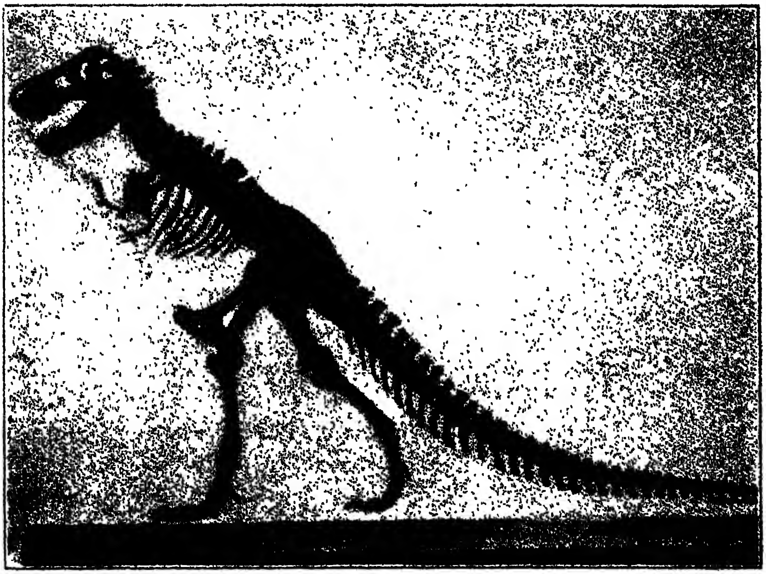


FIG. 3. The skeleton of a great carnivorous reptile (dinosaur) unearthed from Mesozoic rocks, millions of years old. Length, 47 feet. (Courtesy of the American Museum of Natural History.)

each other through time, as the traveler sees them succeed each other in space. At times the drama hastens, and unusual rapidity of geologic action has, in fact, marked those epochs since man has been a spectator upon the earth. (Geological) science demonstrates that mountains are transitory forms, but the eye of man through all his lifetime sees no (important) change, and his reason is appalled at the conception of a duration so vast that the millenniums of human history have not accomplished the shifting of even one of the fleeting views which blend into the moving picture" (J. Barrell).

The known history of the earth has been more or less definitely divided into great eras, and these in turn into periods and epochs. In

he ...tly into the crumbling by weathering; streams have been sawing incessa... ...ds have lands; the sea has been eating into continental masses; the w... life and been sculpturing desert lands; ...usly lo more locally and intermittently...s. glaciers have plowed through mountain valleys, and even vast sheets of ice have spread over considerable portions of continents. The outer shell, or crust, of the earth has shown marked instability throughout geologic time. Slow upward and downward movements of the lands relative to sea level have been very common, in many cases amounting to thousands of feet. Various parts of the crust have been, and are being, affected by sudden movements resulting in earthquakes. During the eons of geological time, vast quantities of molten materials have, at intervals, moved not only into the earth's crust but also often out upon the surface. Mountain ranges hav been brought forth and cut do and sometimes rejuvenated.

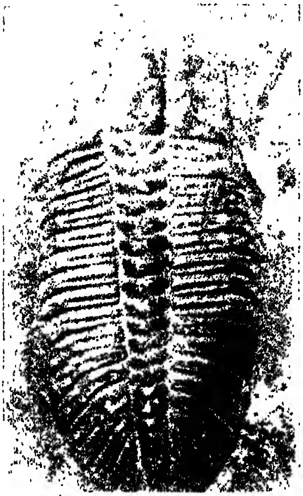


FIG. 2. A fossil sea-animal (Trilobite) hundreds of millions of years old found 9000 feet above sea level in the Rocky Mountains. (Natural size.)

waters have spread over many ...rts of what are today continen...al a (Fig. 1). There have been repeated advances and retreats of the sea over many districts. Lakes have come and gone. Plants and animals have inhabited the earth for many millions of years (Figs. 2 and 3). In earlier known geological time the organisms were comparatively simple and low in the scale of organization. Through the succeeding ages higher and more complex types were gradually evolved until the highly organized forms of the present time, including human beings, were produced.

Vastness of Geological Time. The great importance of the time element in the study of geology cannot be too strongly impressed upon the reader. The length of time of known human history is, indeed, very

the accompanying table, the era and period names, except those representing the earlier times, are mostly world-wide in their usage. Epoch names are too numerous, and usually too local in application, to be included in the table for general use.

TABLE OF MAIN GEOLOGICAL DIVISIONS

Estimated Minimum Duration	*Eras*	*Periods*	*Characteristic Life*	
50 million years.	Cenozoic.	Quaternary. Tertiary.	Age of man. Age of mammals.	Age of highest order of plants and animals.
150 million years.	Mesozoic.	Cretaceous. Jurassic. Triassic.	First high order flowering plants. First birds and modern fishes. First mammals (very primitive).	Age of reptiles and cycad plants.
300 million years.	Paleozoic.	Permian. Pennsylvanian. Mississippian.	Age of amphibians; first insects; and first reptiles. Great coal age (Pennsylvanian) with large, non flowering plants.	
		Devonian.	Age of primitive fishes, and first-known (very primitive) seed-bearing plants and trees.	
		Silurian. Ordovician. Cambrian.	Age of invertebrate animals; first (very primitive) vertebrates in Ordovician; first-known land animals and land plants in the Silurian.	
500 million years.	Proterozoic.	Keweenawan Huronian	Meager records of relatively simple invertebrate animals, and very simple plants.	
500 million years.	Archeozoic.	Timiskaming Keewatin	Primitive life; no determinable fossils.	

Branches of Geological Science

It has already been stated that geology is very broad in its scope, and so with this comprehensive science, as with most other great departments of human knowledge, a number of more or less separate branches or subdivisions have come into existence as special fields of study. Most of the principal branches are as follows:

Mineralogy is the study of minerals, which are naturally formed, inorganic, homogeneous substances of definite (chemical) composition. With the exception of a relatively very slight amount of organic material, minerals constitute the whole lithosphere as far as it is known.

Petrology is the study of the nature and origin of rocks, which are

more or less extensive, solid portions (or formations) of the earth's crust, and which are nearly always made up of aggregates of different minerals, or, more rarely, of masses of a single mineral.

Dynamical geology is the study of the forces (agents and processes) whereby the outer portion of the earth has been, and is being, modified.

Physiography, sometimes called physical geography, in its broadest sense includes a study of the lands, the atmosphere, and the sea. In the United States there is a strong tendency to restrict the field of physiography to the study of land forms, thus making it essentially the same as *geomorphology.* It is concerned mainly with the description, explanation, and classification of the relief features (land forms) of the earth.

Oceanography is the study of the oceans.

Meteorology is the study of the atmosphere. Its chief concern is the weather. The science has grown rapidly in significance with the air-age.

Structural geology is the study of the arrangement, attitude, and relative positions of the rock masses of the earth's crust. Thus, its concern is the architecture of the earth.

Economic geology is the practical application of geology to the arts and industries. It deals with geological products of value to mankind, such as coal, petroleum, ores, building stones, salt, gypsum, etc.

Geophysics deals with the physical forces operating within the earth as a whole. It is a connecting link between geology and physics.

Seismology is the study of earthquakes. It is closely related to both geophysics and dynamical geology.

Paleontology deals with the plant and animal life of the geological ages as revealed by the fossil remains found in the rocks.

Stratigraphy deals with the succession and interrelations of the strata of the earth's crust.

Paleogeography deals with the geographic conditions of the earth during former (geologic) ages, especially with the relations of lands and seas.

Paleontology, stratigraphy, and paleogeography are subdivisions of *historical geology,* which, as already defined, deals with the successive events of earth history, including the history of organisms. The other branches listed above fall in the general area of *physical geology.*

The Concern of Physical Geology

From foregoing statements and from the analysis of the special branches of geological science, one sees that the main concern of the general field of physical geology is, broadly viewed, twofold: (1) the nature and arrangement of the materials which comprise the earth's crust and interior, and (2) the forces or processes, and the various agencies causing them, which operate upon the surface and within the interior of the earth to bring about changes and to produce the earth features which we see today. These items are treated in detail in the subsequent chapters of this text.

The materials of the earth's crust fall into two principal categories, namely, (1) minerals and (2) rocks. The latter are subdivided into three main groups, on the basis of mode of origin—(a) sedimentary rocks, (b) igneous rocks, and (c) metamorphic rocks.

Natural agents of many sorts, such as running water, wind, heat, gravity, and many others, carry on or assist in a great variety of processes (activities) which bring many changes or results upon both the surface and the interior of the earth. Analytically stated, natural *agents* carry on *processes* which produce certain *results*. In accurate scientific thinking one is careful not to confuse these relationships. The processes are the point of departure for much of the treatment in the realm of physical geology. The many, many geologic processes which act upon the lithosphere may be conveniently classified as follows:

A. *Diastrophism*—movements of the earth's solid crust

B. *Igneous activity* or *vulcanism*[1]—phenomena associated with the behavior and movements of hot, liquid rock within and upon the earth's crust

1. Intrusive igneous activity—that which is confined to the earth's interior (i.e., plutonic action)
2. Extrusive igneous activity—that which operates upon the earth's surface (i.e., volcanic eruptions)

C. *Gradation* (denudation)—the combined processes which operate upon the surface of the earth, tending to reduce its many irregularities to a common level

[1] In this book the term *vulcanism* is not preferred because it so strongly ggests the more limited *volcanic activity*.

1. Degradation—those processes of destruction which tend to wear down the higher parts of the earth's surface
   a. Weathering
   b. Erosion
2. Aggradation—those processes of deposition which operate to fill and raise the lower portions of the earth's surface

FIG. 4. Map of North America, showing its main political and physical divisions.

Diastrophism, igneous activity, and gradation, which are very broad activities embracing many subprocesses, are referred to as the three major

earth processes. The most commonplace, yet very significant, is gradation which is in constant operation around us, ever producing changes in our landscape. The principal *agents* of gradation are: (a) atmosphere, (b) running water on land, (c) subsurface water, (d) moving ice, (e) oceans, (f) lakes, (g) gravity, and, to a small extent, (h) organisms.

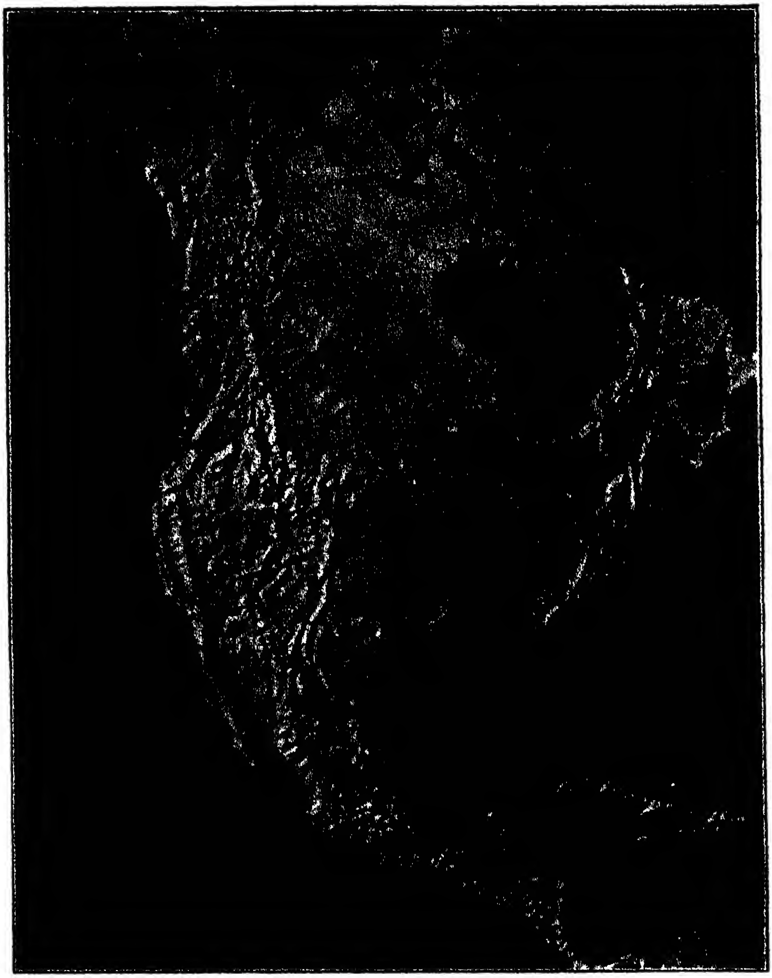


FIG. 5. Relief map of North America. (Courtesy of the United States Geological Survey.)

CHAPTER II

MATERIALS OF THE EARTH'S CRUST—MINERALS[1]

EARTH MATERIALS IN GENERAL

MOST of the hard, naturally formed substance of the earth's crust is referred to as rock. It occurs both in the form of layers (strata) or in irregular masses of various sizes and shapes. It originated in many ways. A small portion came into existence through the deposition and accumulation of organic substances (e.g., coal). Most, however, is made up of aggregates of inorganic particles known as minerals.

Rock, a form of *matter*, is composed of *elements*, which, in turn, are made up of *atoms*. Only 96 different chemical elements are known today, yet they constitute, through various combinations, all the known substances (matter) in existence. Many samples of rock from all parts of the accessible lithosphere have been collected and analyzed so as to give a confident picture of the average chemical composition of the earth's crust. Studies show that of the 96 elements in existence, eight different kinds are sufficiently abundant so as to constitute 98.58 per cent (by weight) of the earth's crust. On the other hand, a considerable number of elements, such as uranium, thorium, and cobalt, are extremely rare in the rocks of the earth. The following list (after Clark and Washington) gives the percentage of crust made up of each of the eight commonest elements:

	Per cent
Oxygen	46.71
Silicon	27.69
Aluminum	8.07
Iron	5.05
Calcium	3.65
Sodium	2.75
Potassium	2.58
Magnesium	2.08

[1] Considerable portions of this chapter are taken by permission from Chapter XXII of the author's *The Story of Our Earth,* which forms Volume 3 of Popular Science Library published by P. F. Collier & Son Company.

All other elements make up but 1.42 per cent of the earth's crust. The table reveals the interesting fact that the two elements in greatest abundance are nonmetallic (oxygen and silicon) and comprise nearly three fourths of the total composition of the crust. The other six elements are true metals. The commonly known valuable metals, gold, silver, nickel, copper, lead, and zinc, which are missing in the table, are rare with respect to the crustal rock as a whole.

Nature and Significance of Minerals

When one examines rocks he finds that most of them consist of various kinds of constituents which can be recognized and identified by certain, specific characteristics. Most such constituents belong in the category called minerals. One may define a *mineral* as a naturally occurring, inorganic substance having fairly definite chemical and physical properties. Thus, to be classed as a mineral, a substance must be of natural origin, must *not* be organic (product of life), must be homogeneous, and must have a composition so definite that it can be expressed by a chemical formula.

All artificial substances, such as laboratory and furnace products (glass, steel, etc.), are excluded from the category of minerals because they have taken no part in the natural development of the earth. Coal and petroleum are not minerals both because of their variable compositions and their organic origin. A few examples of very common substances which satisfy perfectly the definition of a mineral are quartz, feldspar, mica, calcite, and magnetite. About 1500 mineral species are known. To these, and their varieties, several thousand names have been given. Not more than 40 or 50 of the many minerals are, however, of great geological importance, and of these only a half dozen or so make up 90 per cent or more of the outer or crustal portion of the earth. Only two minerals—water and mercury—exist in liquid form under ordinary conditions. The others are solids.

Chemical Make-up of Minerals. Few, however, of the 96 elements of nature occur separately as minerals, such as gold, copper, or silver. In most cases by far, two or more of the chemical elements are variously combined in such a manner (chemically) as to lose their individual identities. Thus the two vicious substances sodium and chlorine are combined to form the beneficial mineral called halite or common salt (composition, chloride of sodium). Oxygen and silicon (a gas and a solid) may be united to form the very hard, common mineral called

quartz (composition, oxide of silicon). Three elements—calcium, carbon, and oxygen—are united in the common mineral known as calcite (composition, carbonate of lime). Four elements—potassium, aluminum, silicon, and oxygen—are chemically combined in the exceedingly common mineral known as orthoclase feldspar (composition, potassium aluminum silicate). Other minerals are more complicated in composition.

Geological Importance of Minerals. Certain rock formations are made up essentially of but one mineral in the form of numerous individual grains, as for example pure limestone which may consist wholly of calcite (carbonate of lime), or pure sandstone which may contain only grains of quartz (oxide of silicon). Most of the ordinary rocks are, however, made up of two or more minerals mechanically bound together. Thus, in a specimen of granite, several distinct mineral species may be distinguished by the naked eye. These mineral grains are from one to five millimeters across. Most common among them are hard, clear, glassy grains, called quartz; nearly white, hard grains, often with smooth faces, called feldspar; small, silvery-white flakes, called mica; and small, hard, black grains, called magnetite.

It is the business of the mineralogist to learn the characteristics of minerals, how they may be distinguished from each other, how they may be classified, how they are found in nature, how they originate, and what economic value they may have. It is an important part of the business of the geologist to learn what individual minerals combine to form the various kinds of rocks (described in Chapter III), how such rocks originate, what changes they have undergone, and what geological history they record. It is thus clear that mineralogy is an important part of geology, which latter is essentially the science of rocks.

Crystal Forms of Minerals

One of the very remarkable facts about minerals is that most of them have a crystalline structure, i.e., they are built up of definitely arranged tiny particles or groups of particles. These constituent particles may be atoms, ions (electrically charged atoms), or molecules (groups of atoms). In almost all naturally occurring minerals of inorganic origin the constituent particles are atoms or ions, but in crystalline substances of organic origin the particles are generally molecules. Crystalline minerals, under favorable conditions, occur in the form of

more or less well-defined *crystals.* A crystal is a geometric solid bounded by regularly arranged plane faces which are a result of a definite or orderly arrangement of its constituent particles (Figs. 6, 394, and 396).[2]

How do crystals develop such regularity of form? Any solid is considered to be made up of many very tiny (submicroscopic) particles held together by attractive forces. In liquids the particles are atoms or groups of atoms which may more or less freely move upon or roll over one

Fig. 6. A group of quartz crystals. (Courtesy of the American Museum of Natural History.)

another, thus permitting alteration in shape of the mass without disrupting it. During the process of change of a substance from the condition of a liquid to that of a solid, because of lowering of temperature or evaporation, or both, attractive forces pull the particles together into a rigid mass. Under favorable conditions, involving not too rapid change, rigid (solid) masses with regular polyhedral shapes will result. This is because the constituent particles have been systematically grouped or arranged. Such a grouping or building up of particles is called

[2] In common usage the term *crystal* is also loosely applied to an irregularly shaped mineral grain which has the definite internal structure, but which has not developed the regular external form because of environmental conditions such as neighbors crowding it (e.g. quartz crystals in granite).

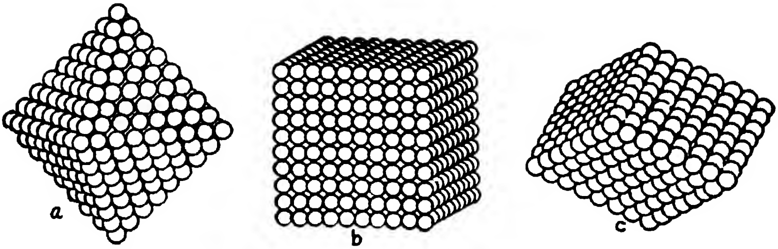


FIG. 7. Piles of shot suggesting the structure of crystals. (After Whitlock, New York State Museum.)

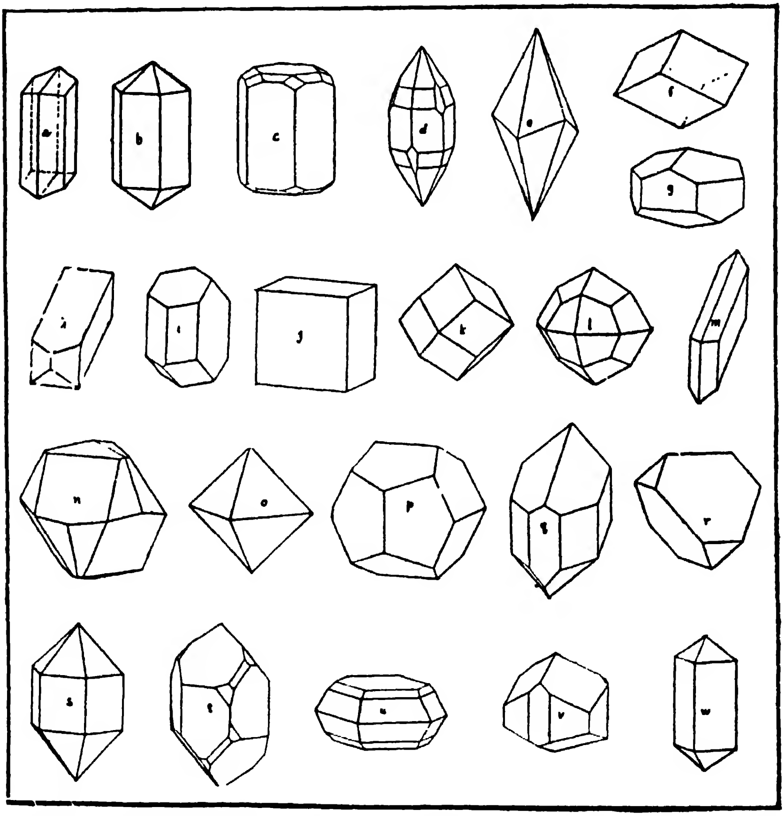


FIG. 8. Crystal forms of some common minerals: *a*, amphibole; *b*, apatite; *c*, beryl; *d*, corundum; *e*, *f*, *g*, calcite; *h*, *i*, feldspar; *j*, fluorite; *k*, *l*, garnet; *m*, gypsum; *n*, hematite; *o*, magnetite; *p*, pyrite; *q*, pyroxene; *r*, chalcopyrite; *s*, *t*, quartz; *u*, sulphur; *v*, tourmaline; *w*, zircon. See Appendix for descriptions. (After New York State Museum.)

crystallization, and the resulting solid is known as a crystal. Crystals grow by addition of material from the outside. They increase in size as long as such material is supplied.

Certain facts furnish all but absolute proof of regularity of arrangement of particles within crystals. Among these facts are the wonderful regularity (*symmetry*) of faces upon crystals; the remarkable property of most crystals to split (cleave) readily in certain directions; the grouping of crystals according to characteristic effects of the passage of light (especially of polarized light) through them; and X-ray photographs showing systematical arrangement of groups of particles within crystals.

All crystals may be grouped into seven systems, each characterized by a certain type of arrangement of crystal faces, angles, and edges about imaginary lines (axes) which run through the center of the crystal. Each system contains from two to seven classes of symmetry, there being thirty-two in all. Each class contains seven fundamental crystal forms. Thus, two of the seven fundamental forms of the class to which garnet belongs are represented by Figs. 8k and 8l. Fig. 8 illustrates perfect crystal forms of a number of common and useful minerals.

Physical Properties of Minerals

Although various means may be employed to identify minerals, it is possible to recognize most of the common minerals by observing a few, simple, physical properties rather than by employing more complicated means of identification such as chemical or microscopic analysis. This is possible since most physical properties are always the same for a given mineral. The following paragraphs describe briefly the more important properties which are useful in the quick recognition of minerals.

Cleavage. Many crystals and crystalline substances exhibit the important property known as *cleavage,* that is, a marked tendency to break or split easily in certain well-defined directions yielding more or less smooth surfaces. A cleavage surface is, as would be expected, always parallel to a possible crystal face, because the splitting occurs along planes of weaker cohesion. The degree of cleavage varies from almost perfect, as in mica, to very poor or none at all, as in quartz. The number of cleavage directions exhibited by common minerals is illustrated by the following: mica, one; feldspar and amphibole, two; calcite and galena, three; fluorite, four; and sphalerite, six. In the study of mineral specimens, careful attention should be given to cleavage whenever it occurs, for certain minerals always show certain cleavage directions.

Hardness. An important criterion for the recognition of minerals is *hardness,* by which is meant the degree of resistance which a smooth mineral surface offers to abrasion or scratching. Scarcely any two minerals are exactly alike in hardness, but for practical purposes a generally adopted scale recognizes ten degrees of hardness as follows:

1. Soft, greasy feel, and easily scratched by the fingernail (e.g., talc).
2. Just scratched by the fingernail (e.g., gypsum).
3. Just scratched by a copper coin (e.g., calcite).
4. Easily scratched by a knife, but does not scratch glass (e.g., fluorite).
5. Just scratches common glass, and is scratched by a knife (e.g., apatite).
6. Not scratched by a knife and scratches common glass easily (e.g., orthoclase feldspar).
7. Much harder than steel, and scratches hard glass easily (e.g., quartz).

8, 9 and 10. Harder than any ordinary substance, and represented in order by topaz, corundum, and diamond.

The ten numbers and the representative mineral-example for each, indicated above, are referred to as the Mohs' *scale of hardness.*

Color. Minerals show a great variety of colors. Many of them, like pure quartz, gypsum, halite, and calcite, are colorless or white. Many of them, like galena (steel-gray), pyrite (brass-yellow), azurite (blue), malachite (green), magnetite (black), and cinnabar (red), possess these colors as inherent characteristics which never fail. Still others, may be variously colored by impurities so that the property is not diagnostic.

Luster. *Luster* is the appearance of the surface of the mineral independent of the color. It is often more or less characteristic of a mineral. The more common lusters are metallic, glassy or vitreous, resinous, greasy, dull, brilliant.

Transparency. A mineral is said to be transparent when an object can be seen clearly through it; translucent when it transmits light but an object cannot be seen through it; and opaque when it transmits no light.

Streak. Certain of the colored minerals exhibit a different color when in powdered form. A simple way to get a sample of the mineral's

powder is to rub the specimen on a piece of unglazed porcelain (so-called "streak-plate"). The *streak* so obtained may be characteristic of the mineral and this greatly aids in identifying the species. Thus, black hematite gives a red streak; black limonite a yellowish brown streak; yellow pyrite a greenish black streak.

Weight. Minerals vary greatly in weight, each one having its own characteristic *specific gravity*, that is, weight in proportion to that of an equal volume of water. The range is from less than 1 to about 23. The average specific gravity of all minerals is about 2.6. It is important to note the relative weight of the specimen examined because it often aids in recognizing the species.

Some Examples of Common Minerals

A number of important characteristics of each of the following common minerals are listed, and the reader should study them with good specimens before him for examination. In this way the properties of minerals in general will be much better understood. Fuller descriptions of these and many other minerals may be found in Appendix A.

Quartz. Composition, silicon dioxide. Crystals, usually six-sided prisms capped by six-sided pyramids. Cleavage, practically none. Colorless or white when pure. Many varieties. Hardness, 7. Specific gravity, 2.6. (See Fig. 6.)

Calcite. Composition, carbonate of calcium. Crystals have faces arranged in sixes or threes around an axis. Cleavage, very good in three directions, none at right angles. Colorless or white when pure. Hardness, 3. Specific gravity, 2.7. (See Fig. 393.)

Mica. Composition, silicate of aluminum with potassium, magnesium, etc. Crystals, usually six-sided plates. Cleavage, excellent in one direction. Colorless, black, brown, etc. Muscovite is a colorless variety, and biotite is black. Hardness, 2 to 2.5. Specific gravity, 2.7 to 3.

Feldspar. Composition, silicate of aluminum and potassium (orthoclase variety) or sodium-calcium (plagioclase variety). The several kinds have common properties as follows: Crystals have prismatic faces meeting at or near 90° or 120°. Color, white, gray, or pink. Cleavage, two good ones at or near 90°. Hardness, about 6. Specific gravity, about 2.6. (See Fig. 394.)

Pyroxene. Composition, complicated silicate of aluminum with various other elements producing several varieties. Crystals, prismatic forms with alternate faces meeting at nearly 90°. Cleavage, two fairly good ones at nearly 90°. Color, usually black, but may be white, green, or brown, according to variety. Hardness, 5 to 6. Specific gravity, 3.2 to 3.6.

Pyrite. Composition, sulphide of iron. Crystals, usually cubes or twelve-faced, each face of the latter being a pentagon. Cleavage, none. Color, brass yellow. Luster, metallic. Hardness, 6. Specific gravity, 5.

Magnetite. Composition, an oxide of iron. Crystals, usually octahedrons. Cleavage, none. Color, black. Luster, metallic. Highly magnetic. Hardness, 6. Specific gravity, 5.

Galena. Composition, sulphide of lead. Crystals, usually cubes. Cleavage, good in three directions at right angles to each other. Color, lead gray. Luster, metallic. Hardness, 2.5. Specific gravity, 7.5. (See Fig. 395.)

Graphite. Composition, carbon. Crystals, usually six-sided plates. Cleavage, good in one direction. Color, black. Luster, metallic. Feel, greasy. Hardness, about 1.5. Specific gravity, 2.2.

CHAPTER III

MATERIALS OF THE EARTH'S CRUST—ROCKS

Introduction

The term *rock,* in a broad sense, refers to any of the solid substance of the lithosphere that can be studied as a unit. So considered, it includes loose, incoherent masses of substance such as gravel, sand, clay, and various forms of soil, as well as firm, consolidated layers and masses such as sandstone and granite. In the strict sense, and as generally used in the classification of rocks, the term is limited to the firm, compacted or consolidated materials. There are many varieties of rocks, formed in many ways. Most rocks are aggregates of minerals; some, on the other hand, are compacted accumulations of organic remains. It is by far most common for a given type of rock to possess two or more mineral species (as in granite), although in a few cases it may consist mainly or wholly of one mineral species (as in rock gypsum, marble, and many limestones). Some rocks consist of five to ten or more different minerals.

Classification of Rocks. All rocks may be classified into three great groups, on the basis of mode of origin or manner in which they were formed, as follows:

1. *Sedimentary rocks*—those derived from sediments. Such sediments, which in due time become compacted, may be the waste product of older rocks which were broken down through the action of various agencies such as atmosphere, running water, glaciers, and others, or precipitates from solution; or they may be accumulations of organic remains. Examples, sandstone, gypsum, and limestone.

2. *Igneous rocks*—those formed by the solidification, upon cooling, of molten rock-material. The sources of such hot liquid rock are within the earth's crust at some unknown depth. The molten substance may flow to the earth's surface before hardening or it may become consolidated at depths below. Examples, lava and granite.

3. *Metamorphic rocks*—those formed by alteration of pre-existing rocks. Either sedimentary or igneous rocks may be so profoundly

changed by geologic agents, such as heat and pressure, as to acquire new characteristics and thereby lose their original identity. Examples, schist, slate, and marble.

General Significance of Rocks. The science of geology is based largely upon the study of rocks, particularly in regard to their origin and history; the forces of nature which affect them; and the events of earth history which they record. It is, therefore, important that the student of geology should gain early at least an elementary knowledge of the more common kinds of rocks. Only by a knowledge of the nature of the materials of the lithosphere (mostly rocks) can the action of geological processes upon them be rightly understood. To this end the student should supplement his reading with study of specimens in the laboratory and of actual rock exposures in the field, as far as that may be feasible.

Sedimentary Rocks

General Characteristics. Rock and mineral matter of any kind carried by water, wind, or ice become *sediment,* which, in the course of time, is deposited as such and later may become consolidated into hard-

Fig. 9. An outcrop showing excellent stratification or bedding. The beds of sandstone (light gray) and thinner (darker) beds of shale have been tilted out of their original horizontal position. Near Oxnard, California.

ened rock. One of the most common characteristics of such sedimentary rocks is their division into layers, i.e., their *stratification* (Fig. 9). By throwing a quantity of loose rock material, the fragments of which range from very fine to coarse, into standing water, the coarsest material would settle first, and upon it successively finer and finer material. There would be a gradation from the coarsest material at the bottom to the finest at the top. By repeating the process a similar *layer* (or *bed*) would accumulate on top of the first, and the two layers would be separated rather sharply by a stratification or bedding surface, called a *bedding plane.* In water with a current there would be a tendency toward horizontal as well as vertical gradation of the sediment in each layer or bed due to the sorting power of the running water. The term *stratum* (plural, *strata*), strictly speaking, applies to a collection of successive beds or layers of the same sort of rock material, but is very often used in the same sense as *bed* or *layer.* It should be borne in mind that some sedimentary rocks show little or no sign of stratification, as for example in the case of many glacial deposits. Wind deposits are often more or less crudely stratified.

Another common characteristic of many sedimentary rocks is the rounded nature of the fragments and particles which compose them. This roundness is due to the fact that any sharp angles of rock and mineral fragments tend to be worn away by abrasion during transportation by water, wind, or glaciers. This feature stands in sharp contrast with that of the typical igneous rocks which are masses of more or less angular minerals (or crystals).

Sedimentary rocks often contain *fossils* (Fig. 2), i.e., remains or traces of animals and plants, whereas igneous rocks seldom contain them.

Where Sediments Are Deposited. The greatest theater of sedimentation is the sea. The general tendency is, and has been for long ages, for the land waste resulting from disintegration and decay of rocks to be carried into the sea, very largely by rivers. Most of this sediment, amounting to vast quantities each year, is deposited in the shallow water relatively near the land, within 100 to 200 miles of the shore. Shells and remains of various animals and plants, as well as volcanic materials (especially dust), accumulate over vast areas of the sea floor. Large quantities of material worn from the shores by wave action are also deposited in the sea.

Most lake bottoms receive sediments both derived by wave action

and carried in by streams. Mineral matter in solution, such as salt and gypsum, may also be precipitated during evaporation of lake water.

More or less deposition of the tremendous amount of sediment carried by streams takes place along the stream courses, particularly on their flood plains, or where streams emerging from mountains flow out into deserts and dry away.

Vegetable matter, which in many places may be changed into coal, accumulates in swamps, bogs, and some lakes.

Various types of sediments are deposited directly upon dry land. Thus piles of rock fragments derived from cliffs often accumulate at their bases; wind, especially in desert regions, transports and deposits great quantities of dust and sand; mineral-charged waters (springs) emerging from the earth deposit their mineral matter at the surface; and glaciers transport and deposit large amounts of rock waste directly upon the land.

How Sediments Are Consolidated. At the time of their deposition sedimentary rocks were loose, incoherent masses. Thus the familiar rock known as sandstone was once loose sand, and shale was formerly soft mud. What causes the consolidation of sediments? One important factor is *weight* or *downward pressure.* Where strata pile up to thicknesses of many hundreds or even thousands of feet as they commonly do, the weight or downward pressure of the overlying masses tends to squeeze together the fragments and particles of the lower masses of the pile, causing them to become compacted.

Cementation is another important cause of consolidation of sediments. Waters penetrating the earth's crust carry various minerals in solution, and at considerable depths such minerals are deposited in the pores of the loose sediment, causing the whole mass to be tightly bound together. The common cements are silica, calcium carbonate, and iron oxide.

Classification of Sedimentary Rocks. On the basis of mode of origin- of the constituent materials, sedimentary rocks may be classified into three main groups: (1) those of *mechanical,* or fragmental, origin (2) those of *chemical* origin, and (3) those of *organic* origin.

Sedimentary Rocks of Mechanical Origin. Sediments comprising these rocks are the fragments of waste from the disintegration of preexisting rocks of the lands. Through the work of natural agents, such as atmosphere, running water, glaciers, and others, former bedrock masses become destroyed, and the broken-up materials transported to

new sites, ultimately to be consolidated into new sedimentary rock. The commonest *sediments* of such origin, before re-cementation into solid rock, are gravels, sands, silts, and clays. The *sedimentary rocks* resulting from consolidation of these are conglomerate, breccia, sandstone, arkose, and shale. These are known as *clastic* rocks.

A CLASSIFICATION OF COMMON SEDIMENTARY ROCKS

Origin of Deposit	*Name of Sedimentary Rock*	*Principal Kind of Material*
	Breccia	Angular fragments
	Conglomerate	Gravel, with rounded pebbles
Mechanical	Sandstone	Sand grains, usually quartz
	Arkose	Sand grains, quartz and feldspar
	Shale	Silt or clay
	Some limestones including: Travertine Lithographic limestone Oölitic limestone Some dolomite	Precipitated calcium carbonate, sometimes with magnesium carbonate
Chemical	Salt	Sodium chloride
	Gypsum	Calcium sulphate
	Bog iron-ore	Precipitated iron compounds
Organic	Most limestones, including: Coquina Chalk Fossiliferous limestone Some dolomite	Calcium carbonate animal remains
	Coal	Plant remains

Sandstones are consolidated sands, usually composed of the mineral quartz. Sand grains range in diameter from $\frac{1}{16}$ to 2 millimeters. They are generally held together by a cement such as lime, oxide of iron, or oxide of silicon (silica). Sandstones are generally stratified in layers varying from thin to thick (Fig. 10). They vary greatly in color according to the nature of the fragments, cementing material, and impurities which they contain. They are most often white, gray, brown, or red. Sandstones are usually very porous because of the numerous, relatively large spaces between the grains of the rock. There are many more or less impure varieties of sandstone, as for example *calcareous*

FIG. 10. A large outcrop of strongly tilted, thick beds of sandstone. Cajon Canyon in southern California.

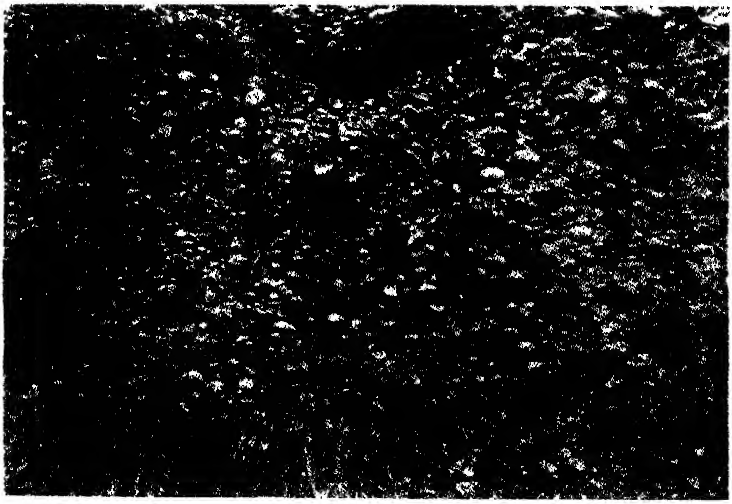


FIG. 11. Detail view of part of a large outcrop of coarse conglomerate. Santa Monica Mountains, California.

sandstone, containing much lime; *micaceous sandstone,* containing many flakes of mica; *argillaceous sandstone,* rich in clay; and *arkosic sandstone,* rich in fragments of feldspar. *Arkose* is a special variety of sandstone containing fragments of feldspar.

Gravels are incoherent masses of more or less rounded pebbles of rock material ranging in size from a few millimeters to boulders a foot or more in diameter. The most common pebbles are of quartz, not only because this mineral is so abundant, but also because it is so hard that the rolling and rubbing action of water, wind, or ice often only rounds off pebbles of it, while softer minerals and rocks are reduced to fine materials. *Conglomerates* are masses of gravel cemented together (Fig. 11). They are usually much more crudely stratified than sandstones because of the conditions under which they are deposited. They are given various names according to the prevailing kinds of pebbles in them, as *quartz conglomerate, granite conglomerate, limestone conglomerate.* If the pebbles, or fragments are angular, instead of rounded as in a conglomerate, the rock is called *breccia.*

Clays consist of very finely divided, decomposed rock or mineral matter of various kinds, but usually mostly of kaolin. The grains are less

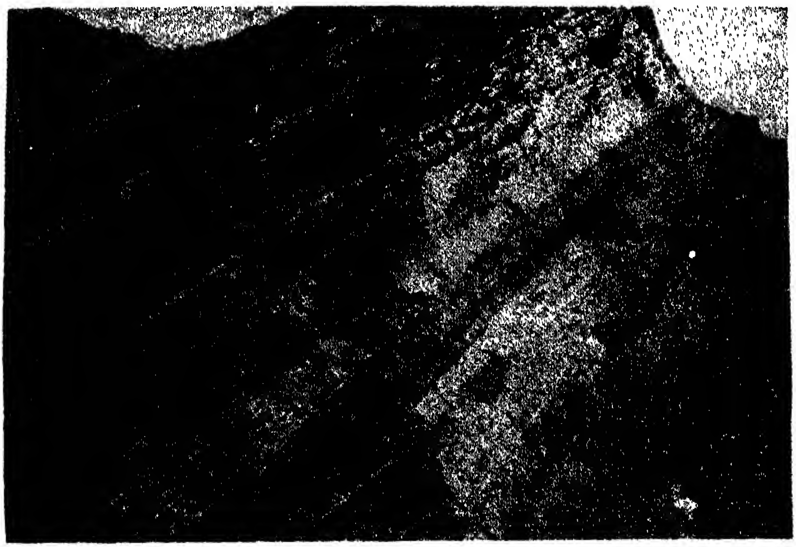


FIG. 12. A large outcrop of strongly tilted, evenly stratified shales. Total thickness of the beds is hundreds of feet Piru Canyon, Los Angeles County, California.

than $1/256$ of a millimeter in diameter. They are plastic when wet. *Silts* consist of nonplastic, finely divided, decomposed or disintegrated rock or mineral matter with grains varying between $1/16$ and $1/256$ of a millimeter in diameter. *Muds* are very finely divided rock or mineral matter mixed with water. *Shales* are consolidated clays, silts, or muds. They commonly vary in color from white through gray to black, bluish-gray to greenish-gray, or yellow through brown to red, the latter colors usually being due to the presence of an iron compound of some kind. Dark gray to black clays and shales usually owe their color to the presence of considerable decomposing organic matter, for example, *carbonaceous shale*. A limy clay is usually called *marl*. There are also *sandy shales*, containing considerable sandy material, and *calcareous shales*, containing more or less limy material. Shales and clays are usually well stratified, often in thin layers (Fig. 12).

Sedimentary Rocks of Chemical Origin. In this category are included all sedimentary rocks whose constituent materials were formed by direct chemical precipitation of mineral matter from solutions. Such precipitates may have accumulated in the seas, in lakes, in streams, or around springs.

Although most *limestones* are composed of calcareous organic remains, a few varieties are formed by the direct precipitation of calcium carbonate in water. *Lithographic* limestone is an extremely fine-grained or dense variety; *oölitic* limestone is composed of miniature, rounded grains of calcium carbonate, resembling fish roe, called oölites; dolomitic limestone, or *dolomite*, contains magnesium carbonate in addition to the usual calcium carbonate.

Salt and *gypsum* are precipitated from salt lakes and lagoons which are subject to excessive evaporation, that is, where evaporation balances or exceeds inflow of water. The tendency is thus for the mineral matter to accumulate in solution until the point of saturation is reached, after which precipitation results. If both salt and gypsum are in solution, the gypsum is deposited first because it is less soluble than the salt. Extensive deposits of both of these minerals exist in many regions, and they are usually well stratified. When pure they are white, but they are often variously colored by impurities.

Travertine and *siliceous sinter* are more or less porous, usually white, spring deposits, being especially conspicuous around the mouths of hot springs. They are both remarkably well developed in Yellowstone Park, the former at Mammoth Hot Springs, and the latter around the geysers.

Travertine consists of limy material which bubbles when touched with hydrochloric acid, and siliceous sinter consists of silica, a compound of silicon and oxygen which is not affected by the acid.

FIG. 13. Detail view of a large mass of travertine. Note the porous, stringy nature of the material. Near Bridgeport, California.

Bog iron-ore is precipitated on the floors of certain bogs or lakes when an iron compound in solution in the water becomes oxidized, and therefore it is insoluble.

Sedimentary Rocks of Organic Origin. Most *limestones* consist of the limy shells, or fragments of shells, or other limy remains of animals and plants, mostly of animals. In many cases the organic remains, or at least fragments of them, are obvious to the naked eye (Fig. 14). In some cases such material can be made out only under a hand lens or microscope. In still other cases the limy organic remains either have been so thoroughly ground up (e.g., by waves on coral beaches) or so completely altered by crystallization that the original organic structures are wholly obscured. Most of the great, very extensive limestone formations have formed on the sea floor by accumulation of limy shells, etc. Limestones are usually well stratified (Fig. 15). *Coquina* is a variety of limestone consisting almost entirely of loosely cemented shells and shell fragments.

Chalk is a very fine-grained, soft limestone with an earthy texture. It is usually white to light gray. Much chalk consists of the limy, microscopic shells of single-celled marine animals. Limestones are often impure because of a mixture with more or less sandy or clayey material, etc., and become *sandy limestones* or *clayey* (or *argillaceous*) *limestones.*

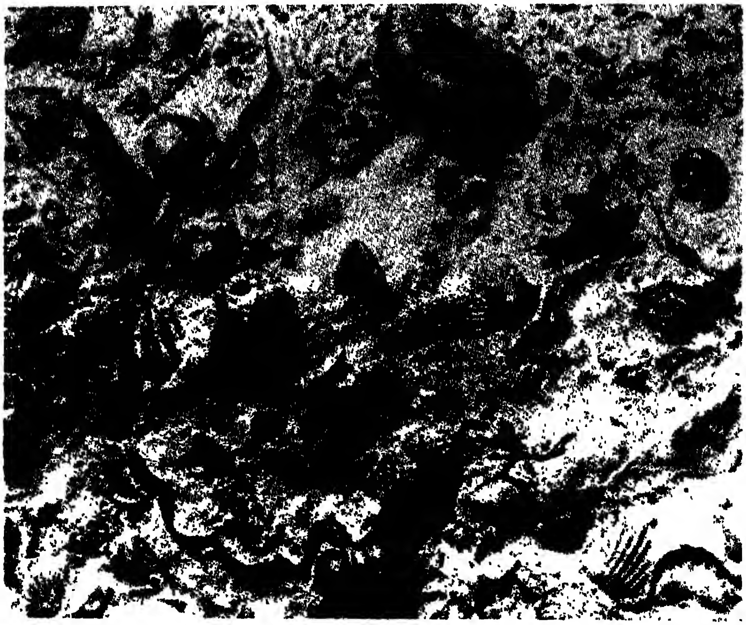


FIG. 14. Part of a specimen of fossiliferous limestone hundreds of millions of years old. It consists largely of shells of various invertebrate animals which once lived in the sea.

Shell marl is clayey material rich in limy shells or fragments of shells. All rocks here described under the category of limestones bubble when treated with ordinary acid. They are usually well stratified.

Diatomaceous earth or *diatomite* is soft, very fine-grained, usually white or gray earthy material composed mainly or wholly of the siliceous tests or secretions of minute, single-celled plants called diatoms. It is usually well stratified, often in exceedingly thin layers. When impure, it is sometimes called diatomaceous shale. It looks much like chalk, but acid does not affect it.

Peat, lignite, and *coal* all represent accumulations of beds of vegetable matter, usually under swamp conditions, which have been more or less decomposed (or carbonized). Vegetable matter of this kind, when only slightly altered, is called peat; when it is somewhat more decomposed it is called lignite (an imperfect coal); and when it is very much changed under conditions of burial in the earth, so that the percentage of

FIG. 15. Well-bedded, tilted limestone hundreds of millions of years old. It contains numerous marine fossils of Ordovician age. Trenton, New York.

carbon is relatively high, it is called coal, including both bituminous and anthracite coal. Coal is more or less well stratified, and it usually forms beds between beds of shale or sandstone (Fig. 376).

Special Features of Strata. *Ripple marks.* These are small parallel ridges, seldom more than a few inches high, formed by the rippling action or current action of either water or wind on certain unconsolidated sediments, especially sands and sandy materials in general. They are particularly characteristic of the action of waves in shallow water. Due to the oscillatory movements of waves, the ripple marks have symmetrical sides and are referred to as *oscillation* ripple marks. On the other hand, those produced by currents are asymmetrical in shape and are called *current ripple* marks. A ripple-marked surface may be hardened enough to be deeply buried under other strata, and later ex-

posed by removal of the overlying strata by the natural process of erosion. Figure 16 shows ripple marks of this kind millions of years old. Ripple marks made by wind action on sand dunes are seen in Figure 288.

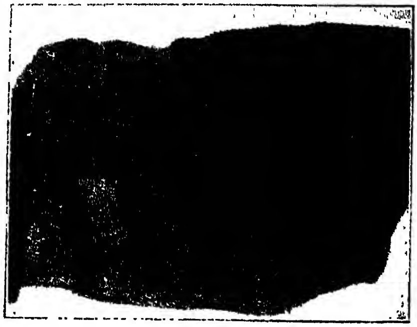


FIG. 16. Ancient ripple-marked sandstone from Holyoke, Massachusetts.

FIG. 17. Mud cracks in a desert, near Cedar City, Utah.

Mud cracks. When soft mud or sandy mud is left exposed to the air after withdrawal of high water, the material dries and cracks into a

network of fissures (Fig. 17). Flood plains of rivers and desert basins, with their alternating wet and dry surfaces, are often very favorable for their development. During dry weather such a cracked surface hardens, and the fissures may either be filled with wind-blown dust or sand, or

Fig. 18. Filled mud cracks many millions of years old on sandstone. Glacier Park, Montana.

the next flood may first fill the cracks and then cover the surface with coarser sediment. Thus a mud-cracked surface may, in the course of time, be deeply buried below the surface and later exposed by wearing away of the land (Fig. 18). *Raindrop impressions* sometimes occur on ripple-marked or mud-cracked surfaces.

Cross-bedding. This is irregular bedding at various angles to the general planes of stratification of a formation (Fig. 19). It is caused by the action of tides or by water or wind currents varying notably in force or direction. Rapid, shifting currents in shallow water of rivers, lakes, and even the sea favor its development. Wind-blown deposits are also often cross-bedded because of the shifting conditions of wind currents carrying sand.

Fossils. Sedimentary rocks often contain remains, impressions, or traces of animals and plants of former geologic ages. Sediments which were deposited in the sea are, as a rule, richest in such fossils (Fig. 2). Lake and river deposits also often contain fossils, as do sometimes the sediments accumulated on land (Fig. 3). Even occasional tracks of land and water animals of millions of years ago are wonderfully preserved.

FIG. 19. Cross-bedded sandstone of Jurassic age. Upper Clear Creek Canyon, tributary to Zion Canyon, Utah. (Photo furnished by the Union Pacific Railroad.)

Concretions or nodules. These are rounded or irregular masses of mineral matter separated sharply and different in kind from the beds or strata in which they occur, and harder than the latter. They are found in a great variety of shapes, sometimes resembling fossil forms. Some small ones are shown in Fig. 20. They range in diameter from less than an inch to a number of feet or yards. They are not pebbles or boulders deposited along with the sediments which contain them as proved by the fact that stratification surfaces often pass right through them. They are segregations, possibly aided by some crystallization, of certain materials, often around some object like a shell or leaf, formed

FIG. 20. Concretions from clay beds of the Connecticut Valley, Massachusetts. One-half natural size.

FIG. 21. Concretions in sandstone. Los Angeles, California.

either during or after the consolidation of the sediment. Their precise mode of origin is, however, not known, although it is believed that most of them developed within the bedrock by the slow accumulation of concentric layers of mineral matter on top of one another. Thus, they "grew" from the inside outward as time went on.

Igneous Rocks

General Characteristics. Igneous rocks are those resulting from the solidification, or hardening, of molten rock-substance originating within the earth's crust. Bodies of such rock are usually massive as compared to the generally stratified character of the sedimentary rocks. In some places, as with lava-flows (Fig. 119), igneous rocks may be piled up in layers, but such rocks can by their other characteristics be told from stratified rocks. Among such other characteristics of igneous rocks are the general uniformity of appearance of masses or layers for considerable distances, both vertically and horizontally; the usual angular, instead of rounded, shapes of many or all of the mineral constituents; the peculiar texture, especially the interlocking of the minerals; their mode of occurrence, especially where they cut across other rocks; the effects of their heat upon adjacent rocks; and their almost utter lack of fossils.

Magma and Its Consolidation. Molten rock-material is called *magma*. Solidification of such magma produces igneous rock. Magma may be regarded as a very hot solution (1500° to 2500° F.) of certain substances dissolved in others. Some of these substances are volatile (gaseous), such as water vapor or steam, carbon dioxide, and sulphurous gases, whereas others are nonvolatile, mainly oxides of silicon, aluminum, iron, calcium, magnesium, potassium, and sodium. These oxides, partly as such, but mostly in numerous combinations involving two or more of them, produce the minerals of igneous rocks. The volatile materials, however, enter but little into the composition of the minerals. The relative amounts of the magmatic substances vary greatly, and hence the igneous rocks produced from magmas show wide differences in both chemical and mineral composition.

When magma under great pressure within the earth moves upward into the earth's crust or to the surface, both pressure and temperature are lowered. Pressure reduction allows the contained gases to escape either into the adjacent rocks within the earth or through volcanoes, often explosively. With a sufficiently slow lowering of temperature, a time comes when some of the nonvolatile constitutents of the magma begin to grow into mineral crystals by systematic arrangement of the unit particles. Finally the whole mass becomes a solid pack of variously oriented crystals, including a number of different kinds (Fig. 22). Because of much interference with each other during their growth, crystal

faces are usually not well defined. Under conditions of very slow cooling, the magma remains highly fluid long enough for the particles to move freely so that fewer crystal centers originate and the rock will, therefore, be comparatively coarse-grained. Very fast cooling causes the magma to become so viscous that the particles cannot arrange themselves into crystals, and so the rock becomes glassy. The cooling may be at such a rate that many local centers of crystallization develop and the rock becomes fine-grained.

Textures of Igneous Rocks. The texture of an igneous rock refers to the shapes, manner of arrangement, and relative sizes of the mineral constituents. From statements in the preceding paragraph one sees that the prime factor causing textural variations is *rate of cooling.*

When most of the mineral grains or crystals making up the rock are of approximately uniform size and readily visible to the naked eye, the rock is said to have a *granitoid texture* (Fig. 22).

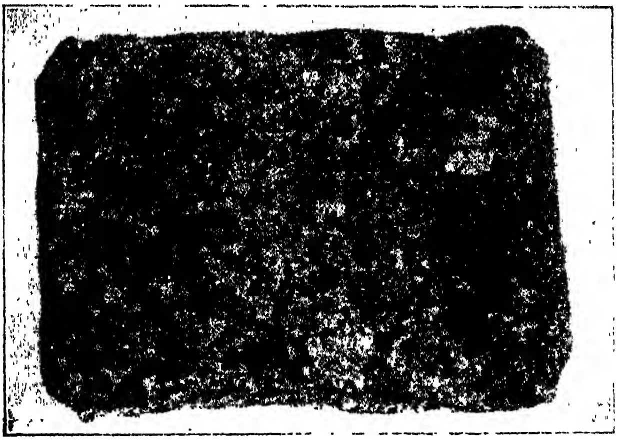


Fig. 22. **A specimen of granite.**

When the rock contains many mineral grains too small to be made out by the naked eye, it is said to have a *felsitic* (or *aphanitic*) *texture.* Such a rock may be partly uncrystallized.

Igneous rocks with relatively large crystals, often with good crystal outlines, which are embedded in a distinctly finer-grained or glassy groundmass, are said to have a *porphyritic texture* (Fig. 23). Such a texture indicates two distinct stages of crystallization.

When a magma cools and solidifies very rapidly with practically no crystallization a *glassy texture* results (Fig. 24). No minerals are formed.

FIG. 23. A specimen of lava showing a porphyritic texture.

FIG. 24. A specimen of obsidian (volcanic glass).

The term *fragmental texture* may be applied to an accumulation (loose or consolidated) of fragments of igneous rocks which have been explosively ejected from volcanoes (Fig. 25).

Scores of mineral species are known to occur in igneous rocks, but

comparatively few of them are abundant. Most common of all are feldspars (both orthoclase and plagioclase), quartz, micas, pyroxenes, amphiboles, magnetite, and olivine. Small amounts of pyrite, apatite, and zircon occur very commonly. Some of these minerals have already been very briefly described, and all them are described at some length in Appendix A.

FIG. 25. An outcrop of volcanic breccia showing a coarse fragmental texture. Ten miles west of Reno, Nevada.

Most or all of the minerals of the fine-grained rocks, and certain of them in coarse-grained rocks, are too small to be recognized by the naked eye or even with the aid of a hand lens. In order to determine such minerals very thin slices of the rocks are made for study with a specially constructed microscope which uses polarized light. By this means all minerals in a rock slice can be recognized because each one has optical properties which distinguish it from all others.

Modes of Occurrence of Igneous Rocks. In the making of igneous rocks, the molten material may become consolidated under two general situations: (a) inside the earth's crust, amid neighboring, pre-existing rock; and (b) on top of the ground where either the liquid rock was poured out as lava flows or solidified igneous fragments were thrown out by volcanic explosions. The first mode of occurrence is referred to as

intrusive or *plutonic,* whereas the second is known as *extrusive* or *volcanic.* A variety of shapes and structures of igneous rock masses exists under both methods of occurrence. Treatment of these is withheld to the end of Chapter V which deals with structures of the earth's crust.

Classification of Igneous Rocks. The principal kinds of igneous rocks may be classified in a general way on the basis of (a) essential mineral contents and (b) texture. In the study of the simplified table presented herewith, it may be clearly understood not only that various accessory minerals are not listed, but also that adjacent types in the classification are not always sharply defined because many intermediate (gradational) types are known.

A CLASSIFICATION OF COMMON IGNEOUS ROCKS

Origin	*Texture*	*Contain orthoclase and quartz*	*Contain orthoclase, but no quartz.*	*Contain plagioclase and biotite or hornblende.*	*Contain plagioclase and pyroxene.*	*No feldspar. Biotite, hornblende, or pyroxene.*
Mainly Volcanic	*Glassy or Fragmental*	Rhyolite obsidian, tuff, breccia.	Trachyte obsidian, tuff, breccia	Andesite obsidian, tuff, breccia.	Basalt obsidian, tuff, breccia.	Limburgite tuff, breccia.
	Felsitic	Rhyolite.	Trachyte.	Andesite.	Basalt.	Limburgite.
Intermediate	*Porphyritic*	Rhyo. and Gra. porphyries.	Trach. and Sy. porphyries.	And and Dior. porphyries.	Bas. and Gab. porphyries.	Lim. and Per. porphyries.
Mainly Plutonic	*Granitoid*	Granite.	Syenite.	Diorite.	Gabbro.	Peridotite.
		Granite family.	Syenite family.	Diorite family.	Gabbro fam ly.	Peridotite family.
		Usually light-colored		Usually dark to black		

Comments on the Classification. The above table involves two important factors—texture and mineral content. Reading horizontally all rocks in a row have a similar texture, and reading vertically all rocks in a column have a similar mineral content. This very brief classification gives a fair idea of some of the more common kinds of igneous rocks and their relationships, but a complete classification would involve many other names and it would be much more complicated.

The chief minerals in the granitoid rocks named in the table are generally large enough to be recognized by the naked eye or with the help of a magnifying glass.

In the porphyritic rocks the relatively large minerals can usually be determined by the unaided eye or with a lens. The other minerals of the porphyries may or may not be recognizable in the same way, depending upon the size of the minerals and the degree of crystallization of the rock. Thus a rhyolite porphyry may be in part uncrystallized or glassy and hence show no minerals or the minerals may be very fine-grained.

Rocks with a felsitic texture are generally so fine-grained, or even partly glassy, that the minerals can seldom be identified without the aid of a microscope.

Glassy rocks of course show no minerals, but, under the microscope, suggestions of some incipient crystals may appear. The place of a glassy rock (obsidian) in the table can be told only by a chemical analysis which indicates about what the minerals would be if the magma had crystallized.

Fragmental rocks can be classified as to mineral content only when enough minerals of the fragments can be made out.

Granite and syenite. A glance at the above table shows that granite is a plutonic rock with a granitoid texture, and that it contains orthoclase feldspar and quartz together with some other minerals in minor amounts. Orthoclase usually predominates. The chief minerals are rather uniform in size and thoroughly interlocked (Fig. 22). Shades of gray and pink are the common colors.

Granite is the most common of all plutonic rocks. It has been intruded in magmatic condition into the earth's crust in large masses in many places and at many times during the earth's history.

Syenite is much like granite, but it has little or no quartz, and it is much less abundant.

Diorite and gabbro. Diorite is a plutonic rock with a granitoid texture containing mainly plagioclase feldspar and dark minerals, such as biotite mica, amphibole (hornblende), and other minor constituents. It is usually dark gray to nearly black.

Gabbro is much like diorite, but it contains pryoxene and often olivine, and it is very dark colored because the dark minerals predominate. It is a common rock (Fig. 26).

Peridotite. This is a plutonic rock with a granitoid texture containing mostly dark minerals, such as biotite, hornblende, pyroxene, and often olivine, but little or no feldspar. It is usually almost black. It is not very common.

Rhyolite and trachyte. Rhyolite is a volcanic rock which has poured out on the surface as lava. It has a felsitic texture and a mineral

composition like that of granite. It is often somewhat porphyritic and ranges to true rhyolite porphyry. The colors are generally shades of gray or pink. Rhyolite is common in some regions.

Trachyte is much like rhyolite, but it has little or no quartz. It is not very common.

Andesite and basalt. Andesite is a volcanic rock in the form of lava. It has a felsitic texture and a mineral composition like that of diorite. The color is usually gray to dark gray. It may be porphyritic.

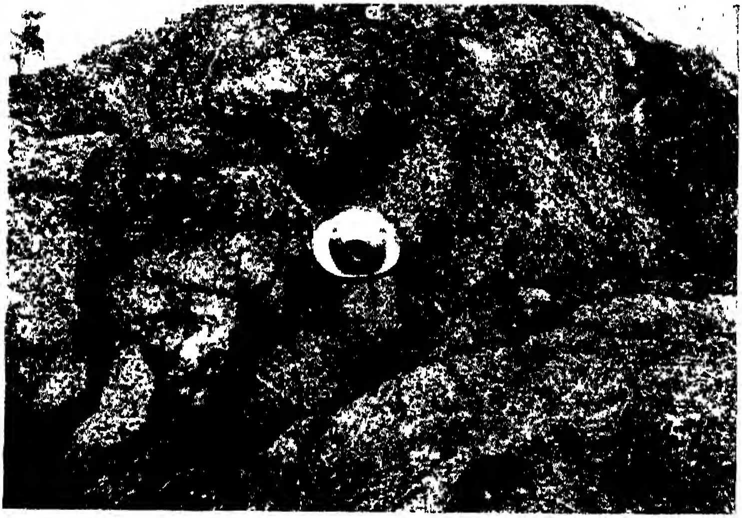


FIG. 26. An outcrop of gabbro which is a plutonic igneous rock. The white material is plagioclase feldspar, and the black minerals are chiefly pyroxene and hornblende. Adirondack Mountains, New York.

Basalt is much like andesite, but it contains pyroxene and often olivine, and it is usually darker.

Andesite and basalt are by far the most common lavas. Successive flows often pile up to thicknesses of hundreds or even thousands of feet (Fig. 119).

Limburgite. This is also a lava. It has a felsitic texture, a nearly black color, and a composition like that of peridotite. It is rare.

Porphyries. The outstanding feature of these rocks is their porphyritic texture. They are designated granite porphyry, andesite porphyry, etc., according to the kind of igneous rock possessing the porphyritic

texture (Fig. 23). They are really textural variants of many igneous rocks. They are very common.

Obsidian. This is also called *volcanic glass.* It has a glassy texture, and it breaks with a shiny shell-like or curving fracture (Fig. 24). It is usually very dark to black and translucent at thin edges. No recognizable minerals occur in it because of the very rapid chilling of its magma. Chemical analysis alone serves to classify it as rhyolite obsidian, andesite obsidian, etc.

Fragmental igneous rocks. These rocks are sometimes called *pyroclastics.* They consist of fragments explosively ejected by volcanoes. Finer-grained material, called *volcanic dust,* gives rise to the rock called *tuff,* while coarser material, of various shapes and sizes, produces *volcanic breccia* (Fig. 25).

Metamorphic Rocks

Meaning of Metamorphism. *Metamorphism,* as applied in geology, refers to any change in mineral composition, structure, or texture of an igneous or a sedimentary rock whereby the original rock character is notably altered. Rocks which have been so modified constitute the *metamorphic* group. In many cases the products of metamorphic action look utterly different from the rocks from which they were derived, whereas in many other cases some of the original features are retained. Simple consolidation of loose sediment, like that of clay into shale, is not regarded as a metamorphic process. Disintegration and decay of rocks under the action of the weather (atmospheric agencies) are, however, processes of metamorphism in the broad sense of the term only.

Sedimentary rocks which have been thoroughly metamorphosed "are much harder, denser, more crystalline, and the fossils, and perhaps even the marks of stratification, have been more or less completely obliterated. As to the igneous rocks, the particular features which distinguish them may disappear, and they may assume a banded appearance and cleavage which resemble those of sedimentary kinds" (Pirsson). Without careful field study, it is sometimes impossible to tell whether a given metamorphic rock was originally igneous or sedimentary.

Agents of Metamorphism. How do igneous and sedimentary rocks become metamorphosed? Brief mention will now be made of the principal agencies of metamorphism. *Liquids,* particularly water on and in the earth, are often effective agents of alteration of rock material by dissolving it, after which it may crystallize in new mineral combinations.

Various *gases* and *vapors,* especially those which escape from molten masses into surrounding rocks, often effect important chemical changes and rearrangements of mineral matter in rocks.

Heat is an important metamorphic agency. By it liquids and gases are rendered much more active chemically. It helps to alter the composition of many minerals and to bring into existence new ones. A sedimentary rock mass may be metamorphosed by heat along the border of a molten mass which is intrusive into the sediment. A rock mass may be heated not only by a hot, near-by igneous body, but also by deep burial within the generally heated crust of the earth. Some heat may also result from disturbances of the crust due to shrinkage of the earth. Heat causes *thermal metamorphism.*

Lateral pressure is a very important agency of metamorphism of both igneous and sedimentary rocks. We shall learn in another chapter that the crust of the earth is, and has been, in many places subjected to tremendous stresses and compressive forces due to earth shrinkage, causing rocks to be bent, mashed, sheared, fractured, and often locally crumpled into mountain ranges. Mineral grains or rock fragments may thus be crushed or flattened; mineral rearrangement may take place; and the rock structure and texture may be greatly changed. Lateral pressure causes *dynamic metamorphism.*

Downward pressure exerted upon deeply buried rocks, particularly sediments, aided by the heat of the earth's interior, and often by water or other liquid, is probably another important factor in the transformation of rocks. This is called *static metamorphism.*

When a rock has its minerals and general composition more or less thoroughly altered by chemical action of hot or cold liquids or gases, it has suffered *chemical metamorphism.* This change may be brought about under conditions of little or great pressure.

In some cases the agencies mentioned may operate only very locally, causing *local metamorphism;* in other cases they may bring about great changes over extensive areas, causing *regional metamorphism.*

A kind of local metamorphism, called *contact metamorphism,* is often brought about by the effect of magma and highly heated gases and vapors emanating from it upon the rocks adjacent to, or in contact with, the magma.

Minerals and Structures of Metamorphic Rocks. Relatively few common minerals make up the great bulk of metamorphic rocks. Important among them are quartz, the feldspars, the micas, the amphiboles, the

pyroxenes, garnet, chlorite, serpentine, calcite, and dolomite. Most igneous and metamorphic rocks are similar in regard to their distinctly crystalline appearance, but in some instances they are unlike in that certain metamorphic rocks have a parallel structure or arrangement of mineral grains, often resembling stratification. In some cases this structure is parallel to original bedding of strata, but more often it is not. On the other hand, some of the metamorphic rocks, such as marble and quartzite, exhibit no parallelism of the component constituents and are massive instead. It should be kept in mind that not all igneous and metamorphic rocks are crystalline. Such cases are, however, relatively exceptional.

FIG. 27. A detail view showing strongly inclined foliation in a mica schist produced by metamorphism of beds of shale. Berkshire Hills, Massachusetts.

Foliation is the arrangement of the mineral constituents of a metamorphic rock with their long axes more or less parallel, thus causing the rock to have a parallel structure. It is a secondary structure, that is, it is developed in the original rock mainly by pressure after the original rock was formed. A rock possessing foliation is called a *foliate.* Well-developed foliation causes *rock cleavage,* that is, a tendency for the rock to split in layers parallel to the foliation. In the usual case of development of foliation some miles within the earth's crust, the mineral grains

arrange themselves with their long axes at right angles to the direction of application of tremendous pressure. The minerals may be old ones forced into such parallelism, or new ones may form by crystallization with their long axes at right angles to the pressure direction. During such development of foliation of a rock, the mineral grains may be more or less crushed or sheared in the near-surface part of the earth's crust, but farther down they are usually deformed without rupture and flattened out at right angles to the pressure direction.

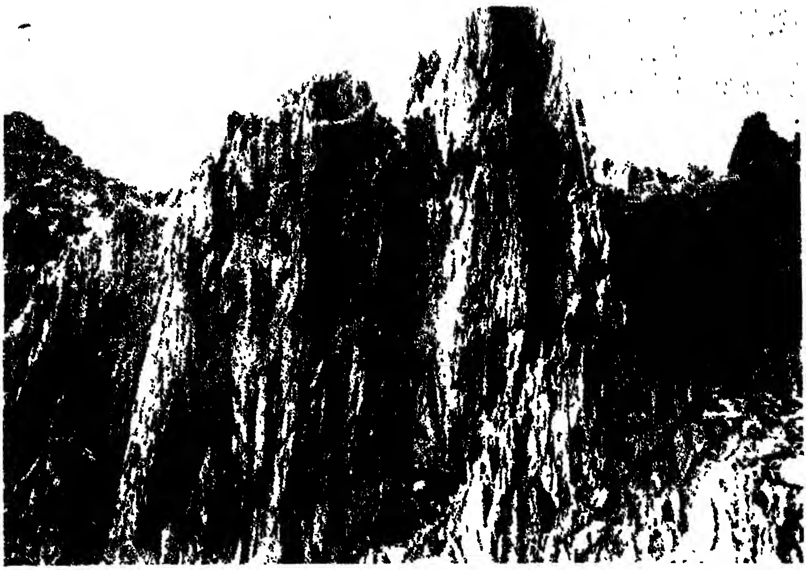


FIG. 28. A detail view showing nearly vertical foliation produced by metamorphism of volcanic dust or tuff. Twelve miles northeast of Victorville, California.

During the production of foliation pre-existing minerals in the rock may, in many cases, be transformed or recrystallized into new mineral combinations, especially where there is sufficient heat and moisture to aid chemical rearrangement of the material.

Former rock structures may also be more or less obliterated by the development of foliation. In some cases the pressure may be applied at right angles to a former structure, such as stratification, and the older and newer (foliation) structures will be parallel or coincide, but far more often the two structures will have different directions. Thus in

the case of strata lying in undisturbed horizontal position, and then foliated by downward (static) pressure the older and newer structures will be parallel. But when such strata have been squeezed out of shape or folded by lateral (dynamic) pressure, a continued application of the pressure or action of a much later pressure may produce foliation (or cleavage) surfaces which cut across the bedding at any angle. Accom-

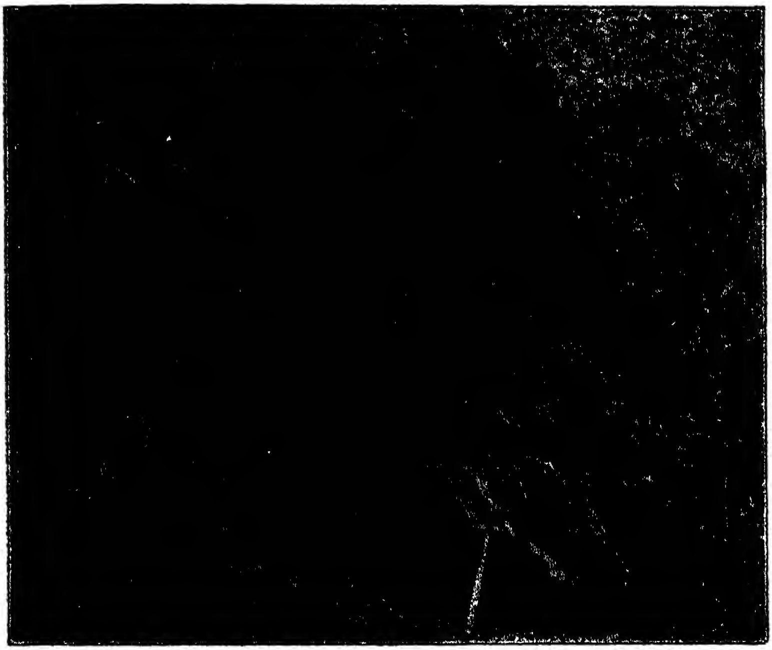


Fig. 29. Slaty cleavage developed across bedding of sharply folded strata. Near Walland, Tennessee. (After Keith, U. S. Geological Survey.)

panying Fig. 29 is an excellent example of stratified rock in which the bent (folded) beds are sharply cut across by the later developed foliation or cleavage. Sufficiently great application of pressure in the case just mentioned would result in obliteration of the bedding surfaces, leaving only a well-developed foliation.

Foliation which is excellently developed in a very fine-grained, not obviously crystalline, rock is called *slaty cleavage*. It is well illustrated by common roofing *slate* (Fig. 391).

A highly crystalline rock with excellent foliation (cleavage), and

platy minerals plainly visible in it, is called a *schist* (Figs. 27 and 28). Both slate and schist split readily in thin, more or less smooth, layers.

A highly crystalline rock with a crudely developed foliation is called a *gneiss* (Fig. 30). Roughly parallel bands of coarse, unlike minerals are developed.

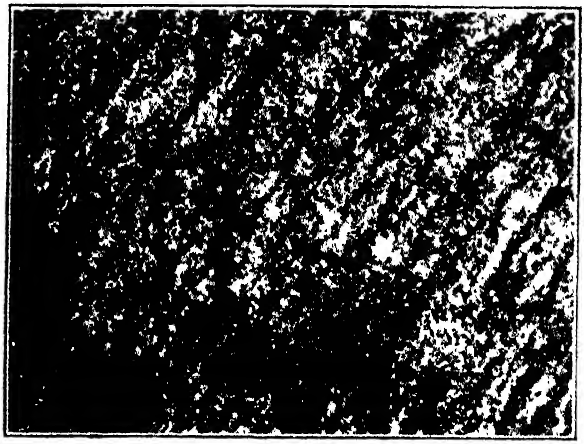


FIG. 30. A specimen showing crude foliation of a granite gneiss.

The term *gneissoid* may be applied as a prefix to the name of an igneous rock in which a crudely developed foliated structure is produced during the consolidation of the magma. In this case the minerals may be crystallizing in a magma while it is being forced upward into the earth's crust under great pressure and the minerals will show a strong tendency to arrange themselves with their long axes parallel to the currents in the still molten material wherever such currents are sufficiently pronounced. The gneissoid structure is, in short, a kind of magmatic flow structure produced under pressure and it is not truly metamorphic.

Classification of Metamorphic Rocks. The metamorphic rocks are, on account of their complicated nature and origin, difficult, if not impossible, to classify satisfactorily. The following table is a grouping of the most important types of metamorphic rocks on the basis of their origin and general structure.

With some brief explanations, rock types listed in the table below may be readily understood. Igneous rocks which have their foliation

impressed upon them after complete solidification of the magma may be named *granite gneiss* or *granite schist,* etc., according to the kind of igneous rock involved. Those whose foliation developed during the process of solidification of the magma may be named *gneissoid granite, gneissoid diorite,* etc., according to the kind of rock. Some *hornblende schist* (or *amphibolite*) is merely foliated hornblende-rich igneous rock. Some *slate* is igneous rock with a typical slaty cleavage. Among the nonfoliated, igneous derived metamorphic rocks, *serpentine* is chemically much altered gabbro or peridotite.

A CLASSIFICATION OF COMMON METAMORPHIC ROCKS

I. Metamorphosed Igneous Rocks	1. Foliates	*a.* Granite gneiss, diorite gneiss. *b.* Gneissoid granite, gneissoid diorite. *c.* Some hornblende schist and gneiss (or amphibolite). *d.* Some slate.
	2. Nonfoliates (massive)	Serpentine, altered lavas.
II. Metamorphosed Sedimentary Rocks	Foliates	*a.* Mica schist and gneiss. *b.* Hornblende schist and gneiss (or amphibolite). *c.* Quartz schist. *d.* Conglomerate gneiss and schist. *e.* Marble gneiss and schist. *f.* Most slate.
	2. Nonfoliates (massive)	*a.* Quartzite. *b.* Marble. *c.* Anthracite coal.

III. Various Gneisses and Schists—some formed by magmatic injection, and some of unknown origin.

IV. Weathered Rocks, Residual Soils, etc.

Many names have been applied to various foliated rocks derived from the sedimentary group. Thus *mica* or *hornblende schist* and *gneiss* are thoroughly crystalline foliates rich in the particular minerals mentioned. Both these minerals may occur in the same rock, usually also with feldspar. They are mainly foliated shales. *Quartz schist* is foliated, impure sandstone often containing mica. *Conglomerate gneiss* is simply foliated conglomerate, and *marble gneiss* is simply foliated im-

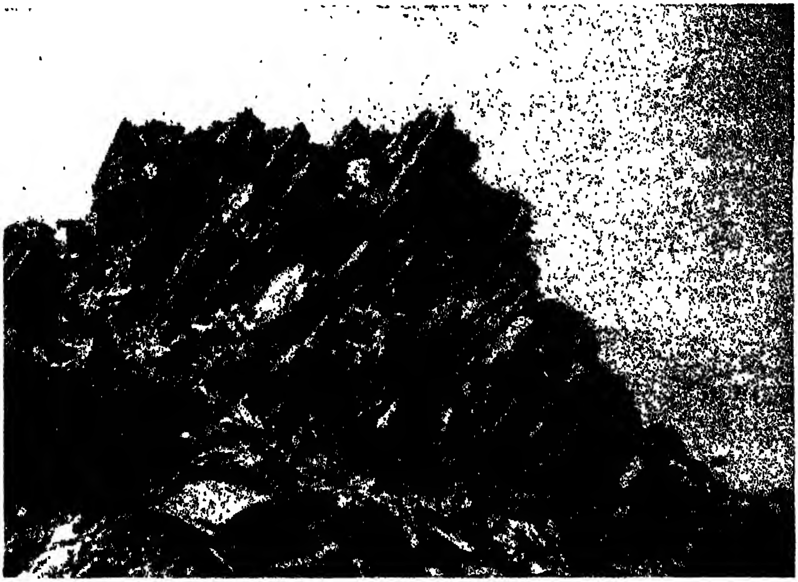


FIG. 31. An outcrop of strongly tilted, alternating, thin beds of marble and quartzite. Maria Mountains, southern California.

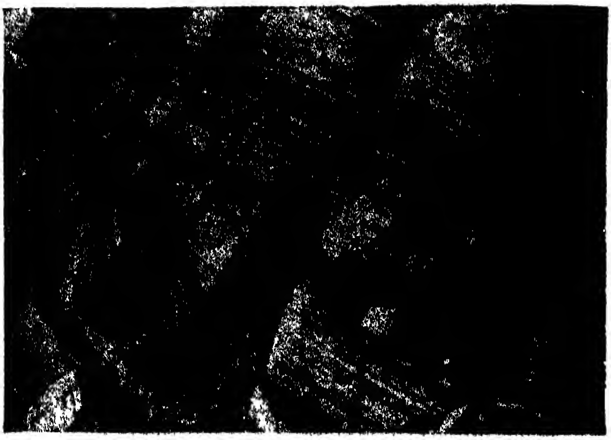


FIG. 32. Banded gneiss. A highly foliated diorite (dark bands) was injected with granite magma in thin sheets parallel to the foliation. Dry Morongo Canyon, southern California.

pure limestone. Most *slate* is foliated shale with highly developed slaty cleavage but with very little crystallization. Among the nonfoliated metamorphosed sedimentary rocks are *quartzite,* a hard, massive rock derived from rather pure quartz-sandstone; *marble,* derived from limestone; and *anthracite,* which is altered bituminous coal.

Some of the various gneisses and schists are formed by more or less intimate penetration or injection of magma into any kind of rock. They may be designated by such terms as quartzite-gabbro injection gneiss, or schist, gabbro-granite injection gneiss or *schist,* etc., according to the rocks involved. They are usually more or less banded (Fig. 32). Various other gneisses and schists may not be definitely known in regard to their igneous or sedimentary origin. They may be designated by rather noncommittal terms such as granitic gneiss or schist, dioritic gneiss or schist, etc., according to their mineral content.

CHAPTER IV

INSTABILITY OF THE EARTH'S CRUST—DIASTROPHISM

INTRODUCTION

Meaning of Diastrophism. The outer shell of the earth is unstable. Overwhelming evidence establishes the fact that it has been so for many millions of years. To the geologist the old notion of a *terra firma* is outworn. The inhabitants of an earthquake country certainly could never have originated the idea of an unshakable, immovable earth. Earth-crust movements may vary from those which are so slow as to be imperceptible to those which are quick and violent. They may be upward or downward or sidewise. They may affect only small, local areas, or they may involve a large portion of either a continent or an ocean basin. The general term *diastrophism* refers to all actual movements of the earth's solid crust of whatever kind or degree.

Types of Diastrophism. In the broadest sense crustal movements fall into two categories, namely, (a) *slow* movements and (b) *sudden* movements, such as earthquakes. Sudden movements are, in the popular mind, more impressive and significant than the slow movements because they are more localized and evident and frequently are accompanied by destruction of life and property, as well as by obvious, though minor, changes in topography. Crustal movements which take place slowly and quietly are, however, often of much greater significance in bringing about profound physical geography changes, such as those which have affected the earth during its eons of recorded history.

In a general way, the slow movements are of two varieties. In one type, known as *epeirogenic* movement, there is either elevation or subsidence of a large or small portion of the earth's crust without notable compression or crumpling (folding) of the rocks, which latter may not have their former attitude changed or they may become gently warped (upward or downward), or more or less tilted. Fracturing and dislocation (faulting) of the rock masses often accompany such an epeirogenic movement which not uncommonly affects a considerable portion of a continent or sea floor. In the other type, known as *orogenic* movement, a relatively long, narrow belt zone of the earth's crust is subjected

to a force of compression, causing the rocks (usually strata) to be more or less crumpled (folded) and upraised into a mountain range.

Various geological agencies, such as atmosphere, winds, streams, glaciers, and the sea, operate externally upon the earth, their general tendency being to cut down the lands and to carry their waste into the sea (degradation). Such agencies would, if not interfered with, completely level the lands and destroy the continents in the course of time. Geological research has made it certain that such external agencies have operated upon the earth for countless ages, and yet the continents have by no means been destroyed. This is because the external agencies are now, and have been throughout recorded earth history, opposed by forces operating from within the earth, that is, by diastrophic forces. Through diastrophism, elevation and re-creation of lands have at least kept general pace with the external forces of destruction; ocean basins have sunk relative to continental areas, causing frequent withdrawals of sea water from areas temporarily submerged; and tremendous volumes of molten materials have been forced not only into the earth's crust, but also out upon its surface. Through sinking of land areas diastrophism has, in many cases, helped to destroy them as such, but, on the average, diastrophic forces which raise up lands (relative to sea level) have predominated over forces which have lowered them.

Regarding the time of crustal movements there is evidence of those movements which have been continuing slowly from late geologic or historic time into the present, in addition to the sudden earthquake movements of today. There is also evidence of crustal movements in the geologic past which have long since ceased and which are recorded in the posture of the rock layers in the form of so-called structures, such as folds, warps, and faults. These structures, bearing testimony of ancient diastrophism, are treated in Chapter V, "Structure of the Earth's Crust."

Slow Movements in Late Geologic and Recent Time

Datum Surface. In land surveying the *datum* is the point, or horizontal line, or surface from which heights or altitudes of points or places are measured or reckoned. The geologist, for his study of the amount and rate of upward and downward movements of the earth's crust, must have some point, line, or surface as a datum. The sea surface is in general the most satisfactory datum since it maintains an average tidal level (within narrow limits) throughout its vast extent. At the

bottom of each topographic map published by the United States Geological Survey there is a statement that "datum is the mean sea level" which means that all elevations recorded on the map are reckoned from the average tidal level of the sea. It should not, however, be assumed that the sea level is, and always has been, fixed and constant. Not only is it a somewhat warped or irregular surface at any given time, but also it may rise or fall very appreciably. In other words, it is not a perfect datum, as will now be briefly pointed out.

It is well known that the earth is not a sphere but rather a spheroid whose polar diameter is about 27 miles less than its equatorial diameter. Approaching the poles, sea level is, therefore, nearer and nearer the earth's center, and so varies with latitude. It is also a warped surface because near lands, especially where large, high mountain ranges lie close to shore, the surface of the sea is drawn upward and toward the lands by gravitational attraction, and so it is disturbed. In extreme cases such distortion is to be measured by a good many feet, but the amount is exceedingly small as compared to the size of the earth. Transportation of sediment into the sea causes rise of sea level by displacement of the water. Sinking of a portion of the sea bottom causes lowering of sea level. Accumulation of ice through snowfall to form great glaciers, as during the Ice Age, represents, in the main, water withdrawn from the sea, and hence a lowering of sea level, just as melting of such ice raises sea level.

None of the variations of sea level above mentioned ever amount to more than a few hundred feet. When it is realized that such variations are very small as compared to the vast expanses of sea and land; that they take place very gradually; and that changes of level between land and sea are generally much greater and more rapid, it is clear that sea level is, after all, a good datum. The records of earth history reveal the fact that many great and small changes of level between land and sea have taken place. Among the minor changes it is often impossible to tell whether it was sea level or land, or both at the same time, which rose or fell. In such cases, therefore, terms like "uplift" and "subsidence," or "elevation" and "depression," as applied to lands are commonly used by geologists in a relative sense only.

Evidences of Elevation of Land. Only a very few of the thousands of definitely known cases of change of level between land and sea will here be briefly described. The examples are chosen to illustrate the more common principles involved. Some of these movements have taken

place within the last few thousand years of clearly recorded human history, whereas others are much older, being records of the geological past.

There are many authentic instances of moderate uplift of the land which have come under the observation of man. A suddent diastrophic movement, resulting in a terrific earthquake, caused uplift of a part of the coast of Alaska near Yakutat Bay to a maximum of 47 feet in 1899 (Fig. 33).

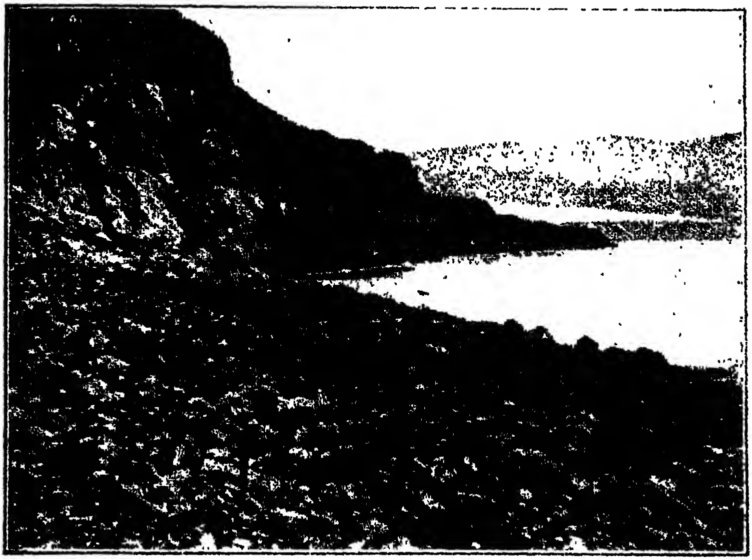


FIG. 33. Part of the shore of Disenchantment Bay, Alaska, which was suddenly uplifted 47 feet at the time of the great earthquake in 1899. (After Tarr and Martin, U. S. Geological Survey.)

Direct measurements by observing marks along the Baltic and Bothnia shores have proved that Sweden has been rising during the last 200 years. At the south the land has remained about stationary, while toward the north there has been increasing elevation, reaching about 7 feet at the north end of the Gulf of Bothnia. Geological evidence indicates that a considerable part of Sweden has been upraised hundreds of feet since the great Ice-Age glacier disappeared from the southern part of the region approximately 13,000 years ago.

"Through slow movements of the crust some of the Japanese harbors are said to have become shallowed one foot in 10 years" (Daly).

Old docks on the island of Crete in the Mediterranean Sea have risen as much as 27 feet within the last 2000 years.

Several rock ledges which were at, or a little below, sea level hundreds of years ago in the Baltic Sea are now distinct islands well above the sea surface.

About 100 years ago a portion of the coast of Chile rose abruptly several feet, causing a severe earthquake.

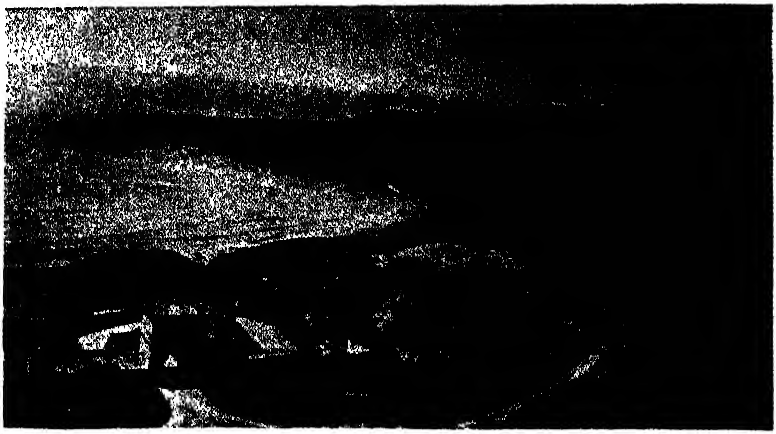


FIG. 34. Part of a great recently uplifted marine terrace facing the sea at an altitude of about 125 feet. The road is on the terrace. Two higher terraces show in profile in the distance. San Pedro Hills, near Los Angeles, California.

Evidence from old elevated shore features, including so-called "raised beaches," is very important. Thus, a succession of terraces cut by the waves of the Pacific Ocean are plainly preserved on the western face of the San Pedro Hills near Los Angeles, California (Fig. 34). The highest and oldest of these terraces is over 1000 feet above the sea, while the lowest, containing many sea shells, is about 100 feet above tide. A somewhat similar succession of terraces occurs on San Clemente Island, about 50 to 60 miles off the southern California Coast. Wave-cut terraces with remnants of rock not removed by the waves occur well above sea level as illustrated by Figure 35. Sea caves formed by wave action are also above sea level in many places (Fig. 36). In Scotland such caves lie fully 100 feet above tide water. Raised beaches and

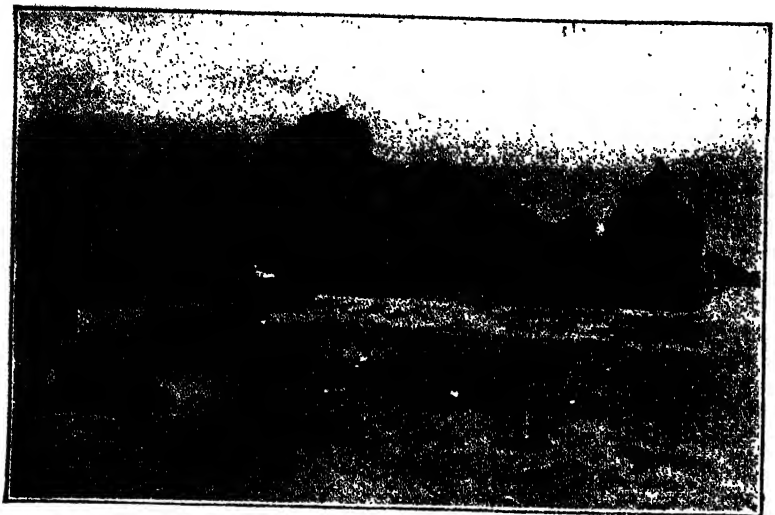


FIG. 35. Elevated marine terrace with remnants of rock which were not cut away by the ocean waves. Near Port San Luis, California. (After G. W. Stose, U. S. Geological Survey.)

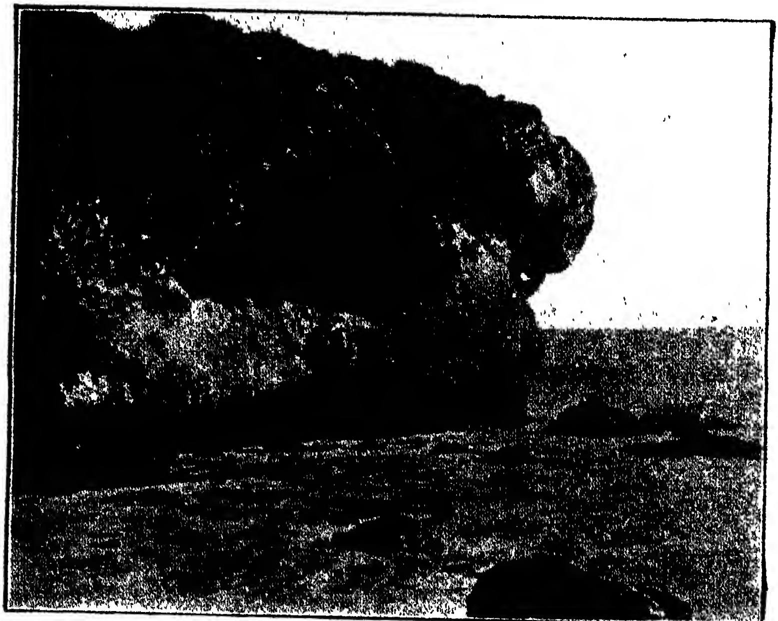


FIG. 36. An elevated sea cave. Near Port San Luis, California. (After G. W. Stose, U. S. Geological Survey.)

shore forms in well-preserved condition up to hundreds of feet above sea level are common in many other parts of the world, for example, Scandinavia, Labrador, west coast of South America, and the West Indies.

Many raised beaches are notably warped or tilted, thus proving that actual earth-crust movements have taken place and not merely a lowering of sea level. A good illustration of this principle is found in the Hudson and Champlain Valleys of New York where marine terrace deposits, formed in the Champlain Sea (Fig. 37), since the great glacier of the Ice Age left the region, lie above sea level. Just south of New York City these raised beaches are only a few feet above sea level, and they gradually increase in altitude to about 600 feet in the vicinity of Montreal. This indicates a differential uplift of the region reaching a maximum of about 600 feet since the glacier withdrew from the region approximately 12,000 to 15,000 years ago.

Remains (fossils) of marine organisms at various altitudes over the world, up to many thousands of feet, afford very strong evidence of uplift of land relative to sea level in late geologic time as well as in the ancient past. There are almost countless numbers of examples. Thus, in the Rocky Mountains of the western United States and Canada sea shells occur in many places at altitudes of from one to over two miles (Fig. 2). The same is true in many other mountain ranges. In Tibet and northern India (Himalayas) fossil marine organisms have been found at altitudes of from three to four miles. In many of these cases the marine fossils are in highly disturbed (folded) strata of geologically recent age. Furthermore, strata of the same geological age lie at all sorts of altitudes in different parts of the world. For these reasons, and in the light of what we have already learned regarding the sea level as a datum, it is evident that such great,

FIG. 37. The Champlain Sea stage of the Great Lakes history when the land stood hundreds of feet lower than now in the St. Lawrence Valley region. (After Taylor and Leverett.)

often differential, changes of level must be diastrophic rather than simply effects of lowering of the sea surface.

Well in the interior of continents, differential earth-crust movements are also known to have taken place. Thus high-level beaches of the vast ancestor of Great Salt Lake have been warped notably. Certain beach lines of ancestors of the Great Lakes have been tilted out of their original horizontal positions to the extent of hundreds of feet, since the Ice Age.

Evidences of Subsidence of Land. In certain parts of Crete old docks have (as already stated) been raised as much as 27 feet above water,

FIG. 38. Post-Glacial marine beds of sand and gravel, containing sea shells, deposited in the Champlain Sea (Fig. 37). They are nearly 500 feet above sea level thus proving at least that amount of uplift since the Ice Age. Eight miles south-southwest of Plattsburg, New York.

while in other portions of the same island remains of similar structures are below sea level, thus proving differential crustal movement.

Portions of the coast of Greenland have sunk recently, as proved by the fact that certain human structures are there below tide water.

Submerged forests prove recent sinking of land in many parts of the world, excellent examples being around the coast of England, especially in Cheshire and Lancashire, and on the shores of the English and Bristol

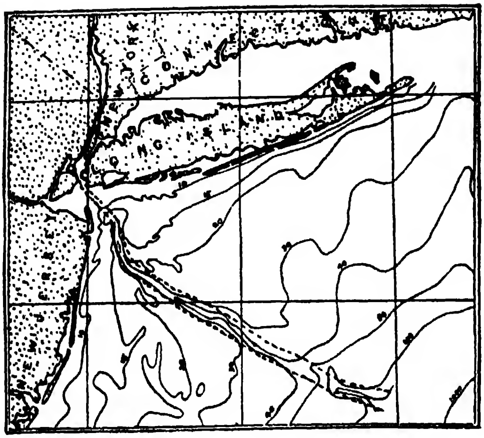


FIG. 39. Map showing the submerged channel of the Hudson River. Figures show depth of water in fathoms. (Data from Coast and Geodetic Survey.)

Channels, where numerous stumps of trees are well below tide-water level.

Certain "parts of Japan have slowly sunk at rates of from one foot in 16 years to one foot in 5 years. Accordingly great rocks formerly visible are now submerged at low tide. Old settlements have been drowned." (Daly.)

A good example of rapid movement of portions of a region in opposite directions at the same time is the Yakutat Bay region of southern Alaska in 1899 where part of the coast suddenly rose as much as 47 feet, while another portion sank below tide level (Fig. 33).

Submerged valleys afford very strong evidence of subsidence, often to the extent of many hundreds of feet.

A fine example of such so-called drowned river valleys is the Hudson Valley of New York. Regional subsidence to the extent of hundreds of feet has allowed tide water to enter this river-cut valley for 150 miles (Fig. 41). Since the Ice Age there has been a partial re-elevation of this subsided region as already pointed out. The Hudson Valley is also very clearly traceable by soundings across the floor of the sea for 100 miles east of New York City (Fig. 39), proving that the earth's crust has there subsided fully 1000 feet since the valley was carved out (eroded) by the river. Notable sinking of the land also has caused a flooding of the lower St. Lawrence Valley by tide water. San Francisco Bay was formed by geologically recent sinking of a portion of the Coast Range region, the Golden Gate marking the submerged channel of the combined Sacramento and San Joaquin Rivers.

In the study of examples of earth-crust movements, it should be clearly understood that a single district may show plain records of both

depression and elevation. In such a case upward or downward movement may be succeeded by movement in the opposite direction. This principle is finely illustrated by the coast of Maine where the whole region sank hundreds of feet in recent geological time, allowing tide

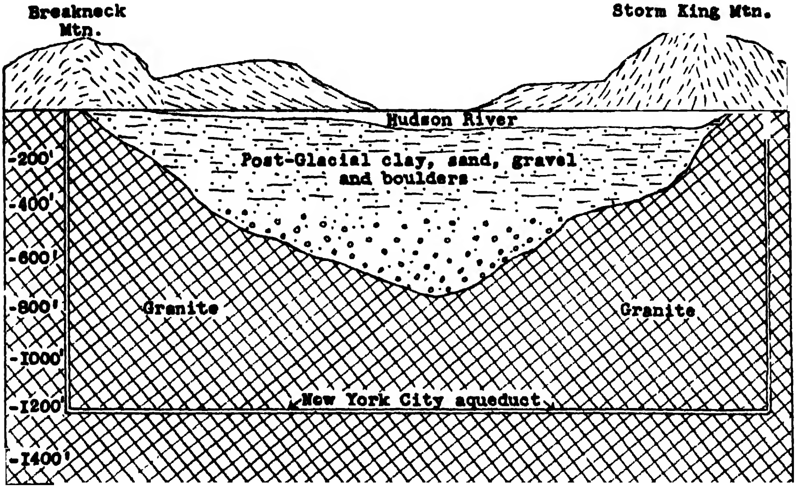


Fig. 40. Structure section across the Hudson River Valley near West Point, New York. The pre-Glacial valley, which has sunk to a depth of nearly 800 feet, is largely filled wtih post-Glacial sediment. The position of the New York Catskill aqueduct is shown. (Redrawn after Berkey, N. Y. State Museum Bull. 146.)

water to flood the mouths and lower valleys of all the rivers, thus giving rise to the deeply indented shore line. A partial re-elevation (of 100 to nearly 200 feet) has taken place as proved by the clay deposits with marine shells along the coast, and for miles up the valleys (Fig. 42).

Cause of Diastrophism

The *fact* of diastrophism is thoroughly established. There is rather general agreement among geologists as to the immediate cause of diastrophism but not in regard to the ultimate cause. The proximate cause appears to be unequal contraction, or shrinkage, of the earth. There is much evidence that the earth, or at least its outer (shell) portion, is heterogeneous, and that it has been shrinking for many millions of years. The fact that strata which, at various times and places, accumulated

under water layer upon layer in horizontal position to thicknesses of many thousands of feet have been highly crumpled and folded into mountain ranges proves earth-crust shortening. In the development of a typical mountain range by this process, the crustal shortening is commonly 5 to 20 miles or more.

FIG. 41. An aerial view of the gorge of the Hudson River in southeastern New York. Because of geologically recent sinking of the region, tide water occupies the valley. (Fairchild Aerial Surveys.)

A general conception is that, as the earth shrinks, its outer shell or crustal portion is subjected to great and small stresses and strains which are relieved occasionally by crumpling of zones of relatively weak rocks, usually strata. In other cases land areas may move upward or downward without crumpling, and with or without tilting. If the earth is a shrinking body, its whole surface must be undergoing a general downward movement toward the center. But, since the earth is a heterogeneous body, not all portions move downward at the same rate, and so the portions which move down less rapidly tend to stand out in relief, giving the appearance of uplift, although their actual movement is also downward.

Viewed very broadly, the earth may be divided into four segments —two oceanic (Atlantic and Pacific), and two continental (Eurasia-Africa and the Americas). It has been proved by actual test (gravity determination) that the materials of the oceanic segments are heavier than those of the continental segments. The oceanic segments are probably moving toward the earth's center faster than the continental segments. Such differential movements would be particularly effective in producing earth-crust disturbances along the general borders of the shifting segments. At the same time the great earth-segments are being more or less divided or broken up into smaller masses, some of which may be subjected to pressure in such manner as to cause localized actual uplifts, as in the folding and uplift of many mountain ranges.

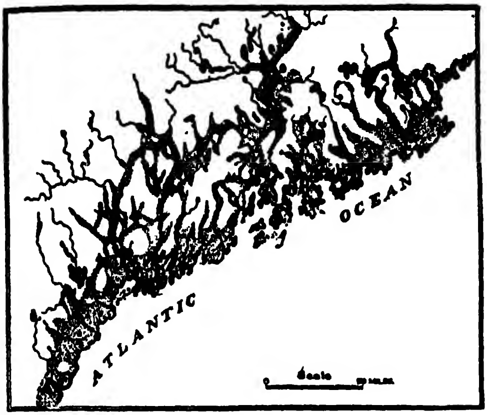


FIG. 42. Sketch map showing the recently sunken and partly re-elevated coast of Maine. Shaded parts indicate marine deposits. (After Stone, U. S. Geological Survey.)

It should be made clear, however, that the ultimate cause of diastrophism is, in our present state of knowledge, far from definitely known —that is, we do not surely know why the earth contracts. It may be due to loss of heat, condensation of earth matter by force of gravity, changes in atomic constitution, or other causes.

Earthquakes

Any sudden movement of a portion of the earth's crust, due to a natural cause, which produces a shaking or trembling of the ground is called an *earthquake*. The study of earthquakes is known as *seismology*. The impulse or shock which gives rise to the trembling originates at a greater or less depth below the earth's surface. Such shocks are known to originate in various ways.

Causes of Earthquakes. Studies during the last fifty years have made it plain that the principal cause of earthquake shocks is the sudden

slipping of portions of the earth's crust past each other along fractures, known as faults. The sudden shifting furnishes the impulse which sends out the vibrations or waves into the surrounding portions of the earth. The first great movement is usually followed for days, or even months, by a succession of after-shocks which generally decrease in number and intensity, although occasionally one or more of the earlier after-shocks may be very severe. Much evidence has been presented recently to support the view that the sudden shifting of the rock along a fault is

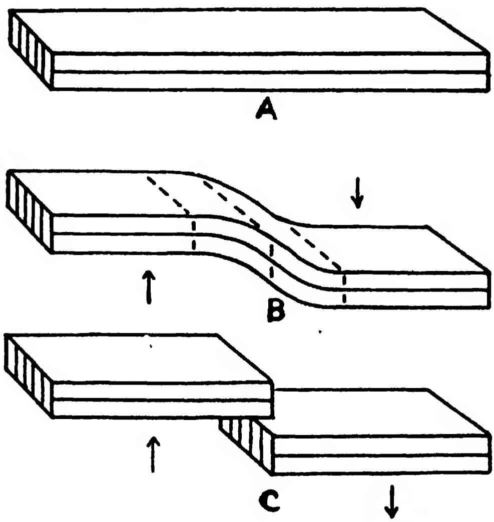


FIG. 43. Blocks of the earth's crust illustrating the elastic rebound theory. A is unaffected by stresses. B is bent by vertical stresses. C shows relief of the strain by faulting.

the result of quick relief of elastic strain which accumulates by slow shifting of adjacent portions of the earth's crust in opposite directions until the rocks can no longer withstand the strain. Such an earthquake may, therefore, be regarded as simply a comparatively sudden manifestation of much slower and broader diastrophic activity.

Recent studies of earthquake waves suggest that some quakes originate at depths as great as several hundred miles where pressure and temperature conditions are very high; that such quakes occur only in certain regions where normal or shallow seismic activity is unusually intense; and that they also are produced by sudden fracture of rocks

where, miles below the surface, great amounts of energy have been storing elastically over long periods.

For many years before the California earthquake of 1906, the crust of the earth underwent increasing strain by geological forces which tended to shear the middle-western part of the state into two parts along the great fault known as the San Andreas rift (Fig. 45). The land near the fault on the western side tended to slip northwestward while that on the eastern side tended to move southeastward. Finally, because

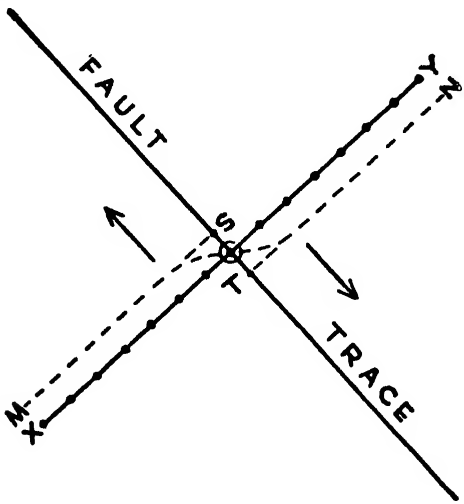


FIG. 44. Groundplan illustrating the elastic rebound theory. XY represents a line (e.g. a fence) crossing a fault before earth-crust stresses have affected the line. MON shows the line bent close to the fault by horizontal stresses acting in opposite directions as indicated by the arrows. MS and TN show the offset condition of the line after the strain has been quickly relieved by fault slippage.

the earth's crust could no longer resist the increasing strain, there was a sudden dislocation along more than 200 miles of the great rift, thus temporarily relieving the strain. In this case the horizontal displacement reached a maximum of 22 feet a little northwest of San Francisco, and diminished to zero in opposite directions along the fault.

According to the elastic rebound theory actual shifting of the land, at the time of an earthquake, occurs only in a narrow zone close to the fault on each side. This theory is now usually accepted as an explanation of the fault type of earthquake in general. The principle involved may be made clearer by inspection of two accompanying figures.

Fig. 43 shows how a block of the earth's crust (A) may be bent (B) by stresses operating vertically until its elastic limit is reached when it suddenly yields by fracture displacement or faulting (C). Fig. 44 is a map or groundplan showing how a portion of the earth's crust subjected to horizontal stresses in opposite directions may yield by sudden fault slippage, allowing the bent portions to straighten out again.

In an earthquake of the kind just explained, the main zone of shock, and, therefore, of destruction, is linear because the vibrations originate in the line of fracture. A very severe earthquake may be caused by a sudden slipping of ten to forty feet along a line of fracture fifty to several hundred miles long.

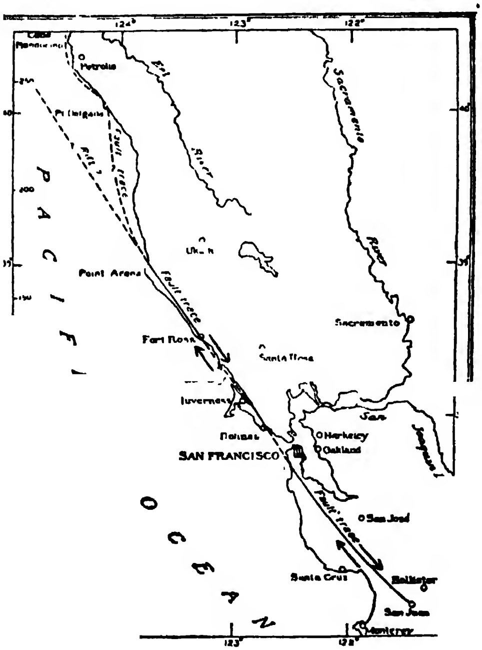


FIG. 45. Map showing the trace of part of the great San Andreas fault, along which sudden slipping caused the California earthquake of 1906. (After U. S. Geological Survey.)

Another, though much less important, cause of earthquakes is volcanic activity. A violent or explosive eruption often causes the earth in its vicinity to quake. Earthquakes not uncommonly immediately precede volcanic eruptions. In still other cases shocks often occur unaccompanied by eruptions in volcanic regions. It is generally believed that such earthquakes are caused by sudden, subterranean yielding of the earth's crust under the influence either of increasing pressure of volcanic gases or of shifting positions of molten rocks imprisoned within the earth and struggling to escape. Earthquakes of volcanic origin are, as a rule, much less severe and more limited in extent than those caused by fracturing of the earth's crust. In volcanic earthquakes the impulse or shock is centralized, rather than linear as in the fracture

type of earthquake, and so the vibrations radiate from the center of disturbance into the surrounding region.

"Very often earthquakes and volcanoes are associated both in space and time. But the ordinary earthquake is not the effect of volcanic action. Both quake and volcano are effects of a common cause, the sudden fracture of a strained crust. In 1914 southern Japan was shaken, nearly simultaneously, at two points. From them seismic waves spread out. These shocks were caused by breaking of the crust. The fracturing changed the pressure conditions of the lava underlying the volcanic cone of Sakurajima. The cone, therefore, erupted violently. Here, then, we have the special case of eruption and shocks, both developed through displacement of the solid crust. Both were due to a common cause." (Daly.)

Minor causes of earthquakes are sudden landslides, submarine slides, collapse of cavern roofs, and falling of rock blocks from brinks of waterfalls.

In many regions neighboring blocks of the earth's outer shell are in a condition of rather delicate equilibrium. Such was the case in the vicinity of Lake Mead before the building of Hoover Dam. The great weight of the lake water, amounting to tens of billions of tons, has caused certain underlying crustal (fault) blocks to be depressed several inches since the peak load of the lake was reached in 1936. This disturbance caused renewed activity along various faults which had been relatively quiet for a long time. Accompanying this activity, several thousand small earthquakes have been recorded in the area.

Frequency, Duration, and Extent of Shocks. Earthquakes are exceedingly common. It is probably true that the surface of the earth is at no given time entirely free from earthquake vibrations. Earthquake recording stations in many parts of the world bear out this statement. Fully 30,000 earthquakes recognizable by the senses occur each year. A great many of these shocks are of course very slight. Only occasionally are the shocks very severe. Earthquakes which cause considerable loss of life and property occur, on the average, perhaps not more than once or twice a year. Earthquakes of sufficient intensity to be felt by people have been recorded in Japan at the rate of several per day, and in Chile, Italy, and California at the rate of several per month, for many years, but most of them have been of low intensity, causing little or no damage.

Among the more damaging earthquakes in California during the last

80 years were: San Francisco region, 1868 and 1906; Inglewood, 1920; Santa Barbara, 1925; Long Beach, 1933; and Imperial Valley, 1940. Severe earthquakes occurring in eastern California, 1872; northern Nevada, 1915; western Montana, 1925; southern Nevada, 1933; and northern Utah, 1934, did little damage because of the sparsely settled regions involved.

The idea that certain regions are immune from earthquakes should not be taken too seriously. Thus the eastern United States is often considered to be immune, but the severe earthquakes at New Madrid, Missouri, in 1811-12; at Charleston, South Carolina, in 1886; and off Nova Scotia in 1929, should dispel this illusion. In New England, which is a region generally regarded as exempt from earthquakes, hundreds of shocks have been recorded within the last 300 years. Probably all but one (eastern New England, 1755) of these have been slight shocks which have caused little or no destruction. Distinct shocks occurred early in 1925. It is true, however, that in certain regions, such as the western United States, earthquakes are much more common than in others, such as the eastern United States.

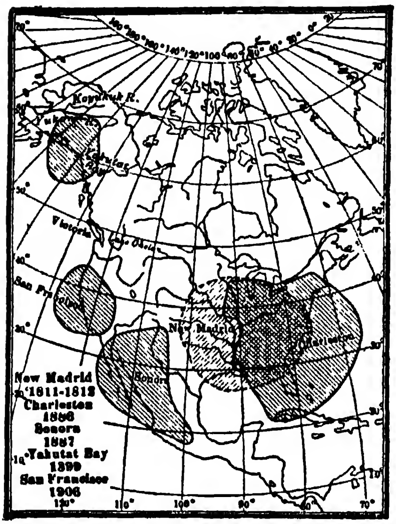


FIG. 46. Map of North America showing areas sensibly affected by some great earthquakes. (From Tarr and Martin's *Physiography*, by permission of the Macmillan Company.)

The vibrations of earthquakes which are sensible to human beings last from a few seconds to several minutes. In general, the greater the intensity of the shock, the longer they last. The average duration of shocks of sufficient intensity to produce much damage is perhaps from one to two minutes.

Earthquakes of sufficient intensity to be noticed by man vary greatly in regard to the size of the region throughout which they may be felt. They may be felt over areas no larger than villages or over considerable portions of continents. The violent California earthquake of 1906 was felt over an area of several hundred thousand square miles. The Charleston, South Carolina, earthquake of 1886 was actually felt by

people over an area of 2,000,000 square miles, and in states as far away as Wisconsin and those of southern New England (Fig. 46). Severe earthquakes, like those just mentioned, actually shake the whole earth, though not enough to be generally recognizable by the senses, as proved by delicate recording instruments in many parts of the world.

Energy of Earthquakes. The energy involved in the more severe earthquakes staggers the imagination. Reid has estimated that the energy of the San Francisco earthquake of 1906 was sufficient to raise a cubic mile of rock 6000 feet, or equal to 800,000,000 times the muzzle energy of a great cannon. But this energy was far less than that of many other well-known earthquakes. Thus, according to Sieberg, the energy of the Japanese earthquake of 1891 was about eleven times at great as that of San Francisco, 1906. Others were still greater in energy.

Someone has estimated that the energy of the San Francisco earthquake was enough to run a battleship constantly full speed ahead for 45,000 years, while that of the Long Beach, California, earthquake in 1933 was only one-one-thousandth as strong.

Nature of Earthquake Waves. In our consideration of earthquakes, the reader should clearly understand that the earth instead of being an excessively rigid body is, as a matter of fact, more or less elastic. A sudden impulse, therefore, sets a portion of the earth in vibratory (or earthquake) motion in somewhat the same manner that a large mass of jelly is set in vibration by a sharp tap on its containing vessel. The vibrations or tremblings travel out in wavelike form into the earth in all directions from the source of the shock. Earthquake waves travel ordinarily at the rate of about three to five miles per second through the outer part of the crust and travel faster with depth.

When, as a result of a sudden shock, vibrations are set up in the earth, as in any solid, they take the form of waves within the earth which are of two important kinds, namely, *longitudinal waves of compression* and *transverse waves of distortion*. In the longitudinal waves, the particles move backward and forward in the direction along which they are transmitted. In the transverse waves, the particles move (vibrate) in a direction across the path of the wave transmission. The longitudinal waves are transmitted at a higher rate of speed than the transverse waves. Furthermore, the transverse waves pass through solids only, while the longitudinal type can pass through either solids or liquids. On reaching the surface of the earth the transverse waves cause the rocking motion of the earthquake. Another kind of wave

travels along the surface, and near surface, portion of the earth. The exact nature of this wave is not known, but in a great earthquake it may throw the ground into a series of actual undulations, somewhat like waves of water, which may be observed to rise in long, low, very swiftly moving waves, causing trees or tall structures to sway violently. The main shock is believed by some to be due to the joint action of transverse and surface waves. At distant points on the earth the kinds of earthquake waves are more or less separately recorded by a delicate instrument called a seismograph, as explained under the next heading.

The actual amount of movement of a particle of the earth's crust during the passage of earthquake waves is generally to be measured only by inches or fractions of an inch. Such to-and-fro motion may, in extreme cases, be six inches to a foot. The shaking is plainly felt, however, when such motion is only a tiny fraction of an inch. At considerable distances from the center of the shock the motion of the earth is exceedingly complex.

Shaking is distinctly more severe in deep soil or loose alluvium, particuarly if it is water-charged, than in bedrock. This explains why buildings some miles farther away from the centeı of disturbance may suffer more damage than those nearer by, as was true in San Francisco in 1906.

Seismograph Records. Instruments of great precision and delicacy, called *seismographs,* have been constructed for the purpose of recording earthquake shocks. The fundamental principles involved are simple, but in actual construction a good seismograph is a complicated machine, a description of which will not here be attempted. In principle, a seismograph involves a heavy mass (say of metal) suspended like a pendulum. On the arrival of an earthquake shock the weight, because of its inertia, remains relatively still for some time, while the earth shakes under it. A marker, such as a pencil, which is attached to the suspended weight also tends to remain quiet during a shock. Now, if a recording plate or cylinder is set in the earth immediately adjacent to, and in contact with, the marker, it is evident that the recording plate or cylinder will move with the earth during a shock and thus be marked by the pencil point. Another weight suspended from a spiral spring keeps its position during up-and-down motions of the earth and so affords a ready means of recording such motions. A seismograph record is called a *seismogram* (Fig. 48). The best seismograms are recorded on rotating cylinders

because on them the lines are not superimposed upon each other. Figure 47 illustrates the seismograph principle.

A good seismograph is practically automatic in its operation. Upon its several rotating cylinders, which are run by very precise clockwork, the north-south, east-west, and up-and-down components of motion; the exact time of beginning and ending (and therefore duration) of the shock; and the intensity of the shock (usually magnified) are all recorded.

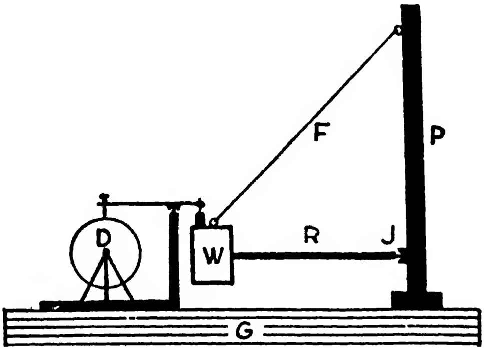


FIG. 47. Diagram showing the principle of a seismograph. G, ground; P, post set in the ground; W, weight; R, rigid support contacting the post with a free-moving sharp point at J; F, flexible wire; D, recording drum revolved by clockwork; and the marker extends from W to D. When the ground shakes the suspended weight, because of its inertia, scarcely moves, but the shaking motion is transmitted to the marker which leaves a record on the drum.

It has been deduced, from a study of seismographic records of distant earthquakes, that two sets of preliminary tremors immediately precede the main shock. The first preliminary tremors (primary waves) seem to be the longitudinal waves, and the second preliminary tremors (secondary waves) seem to be the transverse waves. With the exception noted just beyond, both of these pass through the earth along chords from the place of origin of the shock to the seismograph station. The larger surface waves, which may be combined with transverse waves, pass around the earth in its surface portion in all directions from the seat of disturbance. These waves constitute the main shock (Fig. 48). Preliminary transverse waves do not pass through the earth below a depth of about 1800 miles. This strongly indicates that the inner portion of the earth, over 4000 miles in diameter, is of a different

nature than the outer portion. Since transverse waves pass through solids only, the inner portion of the earth may possibly be liquid.

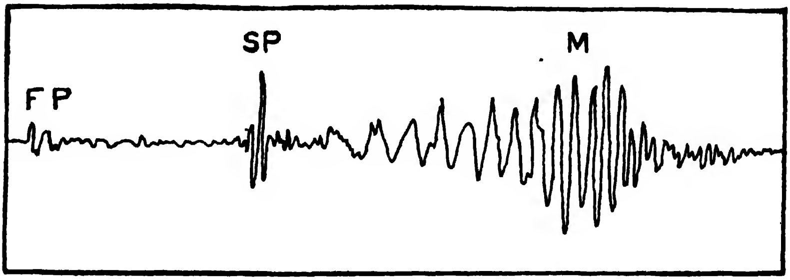


Fig. 48. A seismogram showing the record of an earthquake 2500 miles from its source. FP and SP are the preliminary tremors. The first, FP, called the primary, were produced by fast longitudinal waves which traveled through the earth along a chord; the second, SP, called secondary, were caused by slower transverse waves which followed along the same path. Most of the waves which produced the main shock, M, traveled along the earth's surface.

Intensity Scale. In regard to their magnitude, two general classes of earthquakes may be suggested: (1) those which disturb large sections of continents and actually set the whole earth in slight vibration, and (2) those which affect only local areas with radii of not more than a few hundred miles.

In regard to intensity of individual earthquakes, the modified Mercalli scale has been adopted quite generally. In abridged form it is as follows:

I. Not felt except by a very few under especially favorable circumstances.

II. Felt only by a few persons at rest, especially on upper floors of buildings. Delicately suspended objects may swing.

III. Felt quite noticeably indoors, especially on upper floors of buildings, but many people do not recognize it as an earthquake. Standing motor cars may rock slightly. Vibration like passing truck.

IV. During the day felt indoors by many, outdoors by few. At night some awakened. Dishes, windows, doors disturbed; walls made creaking sound. Standing motor cars rocked noticeably.

V. Felt by nearly everyone; many awakened. Some dishes, windows, etc., broken; a few instances of cracked plaster; unstable objects overturned. Pendulum clocks may stop.

VI. Felt by all; many frightened and run outdoors. Some heavy furniture moved; a few instances of fallen plaster or damaged chimneys. Damage slight.

VII. Everybody runs outdoors. Damage negligible in buildings of good design and construction; considerable in poorly built or badly designed structures; some chimneys broken. Noticed by persons driving motor cars.

VIII. Damage slight in specially designed structures; considerable in ordinary substantial buildings with partial collapse; great in poorly built structures. Panel walls thrown out of frame structures. Fall of chimneys, factory stacks, columns, monuments, walls. Changes in well water.

IX. Damage considerable in specially designed structures; well-designed frame structures thrown out of plumb; great in substantial buildings, with partial collapse. Buildings shifted off foundations. Ground cracked conspicuously. Underground pipes broken.

X. Some well-built wooden structures destroyed; most masonry and frame structures destroyed with foundations; ground badly cracked. Rails bent. Landslides considerable from river banks and steep slopes. Water splashed (slopped) over banks.

XI. Few, if any, (masonry) structures remain standing. Bridges destroyed. Broad fissures in ground. Underground pipe lines completely out of service. Earth slumps and land slips in soft ground. Rails bent greatly.

XII. Damage total. Waves seen on ground surfaces. Lines of sight and level distorted. Objects thrown upward into the air.

Effects of Shocks. Earthquakes are generally classed among the most terrifying of all natural phenomena because of the awful loss of life and property which sometimes results from them. Among the many very destructive earthquakes of modern times mention may be made of several as follows: Lisbon, Portugal, in 1755, when practically the whole city with its population of 60,000 was destroyed; Naples, Italy, in 1788, which cost 32,000 lives; Indus Valley, India, in 1819, which was very destructive of both life and property; Chile in both 1822 and 1835; Assam, India, in 1897, which killed many thousands of people and destroyed much property; California, in 1906, which directly and indirectly (through fire) destroyed much property, but not many lives; Messina, Sicily, in 1908, which destroyed much property and killed approximately 200,000 people; Tokyo, Japan, in 1923, which destroyed much of the city, and also Yokohama, with a loss of life of approximately

150,000; and Kansu, China, in 1920, at least 100,000. Mallet has estimated that at least 13,000,000 people have been killed by earthquakes during the last 4000 years.

Earthquakes may also cause changes in the earth's surface. In this connection it is important to keep in mind *cause* and *effect* of earthquakes. Thus the actual sudden shifting of portions of the earth's crust along either side of the line of fracture (fault), which is often accompanied either by the development of a fissure, or a steep declivity along the line of fracture (Fig. 49), is the cause rather than an effect of an earthquake. Numerous changes are, however, direct effects of shocks, even at considerable distances from the seats of disturbance. Thus, the vibrations often set off landslides, especially in mountainous regions. Cracks and fissures and local small elevations and depressions of the land often occur, and they may affect surface drainage. The disturbance of the earth's crust may cause old springs to stop flowing or new springs to develop. An extraordinary subsidence occurred during the Indian earth-

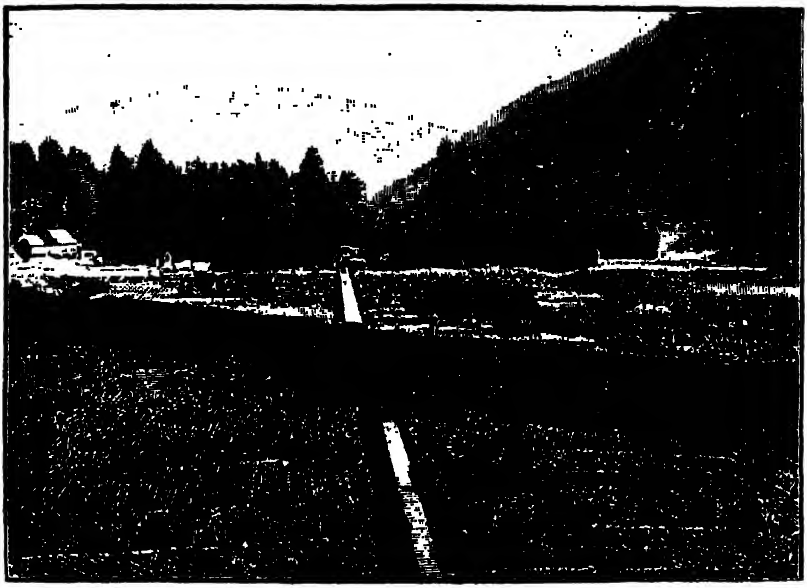


FIG. 49. A fault scarp suddenly formed at the time of the Japanese earthquake of 1891. (After Kôtô.)

quakes of 1819 when a tract of land covering some 2000 square miles near sea level actually sank a little below sea level. It very rarely happens, however, that even a great earthquake produces more than very minor topographic effects.

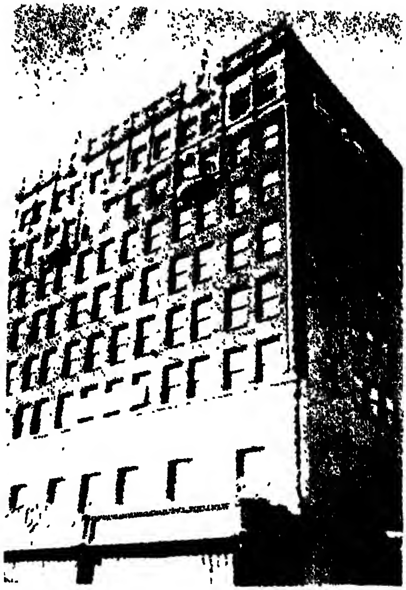


FIG. 50. A tall, well constructed, steel-frame building practically unaffected by the Santa Barbara, California, earthquake of 1925. Compare with Fig. 52.

Earthquakes and Building Construction. Buildings, particularly in regions known to be subject to earthquakes, should be built with earthquake hazard in mind. It is probably more feasible to construct buildings to withstand successfully all but rarely occurring, very severe earthquakes than it is to build them against cyclones or hurricanes. Many studies of earthquake effects upon buildings have made certain lessons very plain.

Well-built wooden buildings, securely anchored to good foundations, are excellent earthquake resisters and very few people are ever killed in such buildings.

Masonry buildings, as often constructed, are easily damaged or wrecked (Fig. 52). On the basis of personal observations after both the Santa Barbara and Long Beach earthquakes in California, the writer is convinced that by far most of the damage to buildings involved those of ordinary masonry construction. Schoolhouses, theaters, churches, and other buildings with large rooms and of improper masonry construction are especially liable to damage or demolition. Masonry buildings well constructed of good materials with cross-walls or enough strong partitions, all securely tied together, are good earthquake resisters. Masonry in the auditorium type of building should be reinforced or tied by steel.

Steel-frame buildings, even skyscrapers, withstand hard shocks surprisingly well (Fig. 50). Reinforced concrete also stands well.

Buildings on made ground or loose alluvium are much more liable to damage than those on solid ground or bedrock.

In all cases careful attention should be given to foundations and the tying of buildings to foundations in such manner that they can with-

Fig. 51. Ground torn up along the San Andreas fault at the time of the California earthquake of 1906. (After G. K. Gilbert, U. S. Geological Survey.)

Fig. 52. Buildings showing sharp contrast in resistance to earthquake action. Building on left unaffected, while flimsy brick walls on right fell away. Santa Barbara, California, 1925.

stand a horizontal stress at least one-tenth as great as the weight of the building.

Dams should not be built on faults known to have been recently active, and preferably not on a fault at all.

Distribution of Earthquakes. Although earthquakes are very widely distributed, so that no part of the earth seems to be immune from at least slight tremors, nevertheless most of them by far occur within *two* great, rather crudely defined belts or zones, as shown by Figs. 53 and 54. One of these belts almost encircles the great Pacific Ocean, and the other extends in a nearly east-west direction around the earth through southern Asia, the Mediterranean district, the Azores, the West Indies, Central America, the Hawaiian Islands, and the East Indies. In a study of 170,000 earthquakes, Montessus de Ballore found that nearly 95 per cent of them occurred within these two belts. It is an illuminating fact that not only the great majority of active and recently active volcanoes but also most of the youngest mountains of the world are located within the two great earthquake belts. In fact it seems rather clear that both earthquakes and active volcanoes are only surface, or near-surface, manifestations of the great diastrophic forces which, at the present time, are operating chiefly within these two belts, but in the present state of our knowledge we cannot say why these great forces are there so active. A study of the ancient records of the earth (historical geology) shows that diastrophism has by no means always been especially vigorous within these two belts.

FIG. 53. Map of the Western hemisphere showing the principal earthquake regions. (After M. de Ballore.)

Submarine Earthquakes and Tsunamis. Many earthquakes are known to take place under the ocean, mostly within the belts just de-

FIG. 54. Map of the eastern hemisphere showing the principal earthquake regions. (After M. de Ballore.)

scribed, but obviously our knowledge concerning them is more meager than it is concerning earthquakes on land. Submarine disturbances are felt on shipboard, and ocean cables are sometimes broken by them. Among very recent, severe, submarine earthquakes, mention may be made of one which took place early in 1946 on the sea floor along part of the southern side of the Aleutian Islands. Resulting waterwaves traveled to shores as far away as the Island of Hawaii where they caused destruction of property. Such giant sea waves, known as *tsunamis,* are caused by the sudden movements of portions of the sea bottom. They are often miscalled "tidal waves."

Tsunamis may be from 50 to 200 miles from crest to crest, and up to 40 feet high, where they originate. They travel with a speed of hundreds of miles per hour, but in the open sea they are scarcely noticeable because they are so broad and relatively low. Tidal gauge records show that certain tsunamis from Japan have crossed the Pacific Ocean, with height diminished to less than a foot, in about 12 hours. If a great tsunami starts reasonably near a coast it will pile up in passing into shallow water, and it may sweep upon the land in the form of a huge surge or breaker from 25 to 100 feet high. Such an earthquake seawave swept over part of the city of Lisbon, Portugal, in 1755 with destructive violence, and another in Chile (1868) carried a United States war vessel half a mile inland, and left it stranded.

Prediction of Earthquakes. Geologists are often asked if earthquakes can be predicted. When and where will the next hard shock occur in a region? The answer is that they may be predicted in a very general way only, that is, in a region which has been notably shaken at least

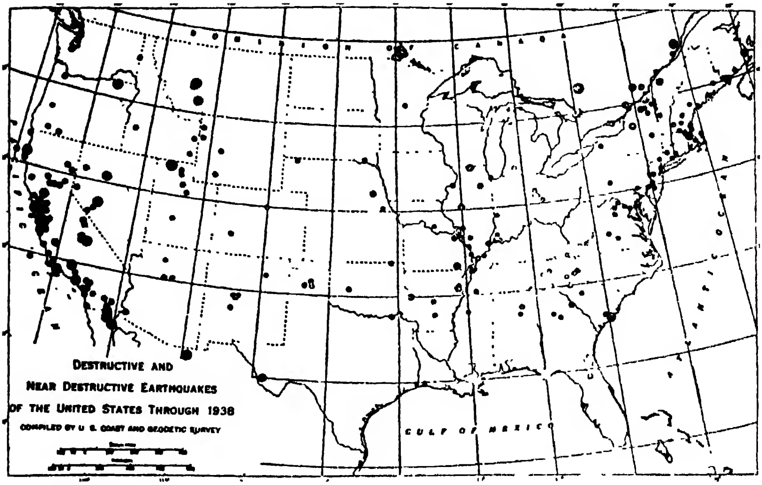


FIG. 55. Map showing the distribution of destructive and near destructive earthquakes in the United States to 1938 inclusive. Most sections of the country have experienced them, some much more than others. The Pacific Coast ranks first, the Atlantic Coast second, the interior of the Mississippi Valley third and the Rocky Mountain region fourth. (Compiled by the U. S. Coast and Geodetic Survey.)

FIG. 56. Buildings wrecked by the Charleston, South Carolina, earthquake of 1886. (After Hillers, U. S. Geological Survey.)

several times within a period of say 50 or 100 years it is almost certain that other shocks will occur within a similar period of time.

If, according to the elastic rebound theory, "earthquakes are due to a straining of the earth, we may be able to watch the progress of the strain and ultimately learn to prophesy where, and perhaps when, the crust will snap. Fairly elaborate plans are already worked out for measuring earth strain which develops as the years go on. It may take a generation or two to accumulate sufficient data for the purpose of forecasting the place, and less probably the time, of earthquakes." (Daly.) If this, or some other method can be worked out, it would, indeed, be a wonderful contribution of science to human welfare.

Typical Examples of Great Earthquakes. Our present purpose is to describe very briefly some selected examples of great modern earthquakes in order better to impress upon the reader many of the more important phenomena which they exhibit.

New Madrid, Missouri, in 1811–1812. The many earthquakes which affected a large district around New Madrid, Missouri, in 1811–1812 were remarkable not only for their great severity, but also because they occurred well outside of the two major seismic belts of the world, and in a region far from volcanoes or growing mountains. The first great shock came during the night of December 16, 1811. "Early in the morning another shock, preceded by a low rumbling and fully as severe as the first, was experienced. The ground rose and fell as earth waves, like the long, low swell of the sea, passed across its surface, tilting the trees until their branches interlocked, and opening the soil in deep cracks as the surface was bent. Landslides swept down the steeper bluffs and hillsides; considerable areas were uplifted; and still larger areas sank and became covered with water emerging from below through fissures or little 'craterlets,' or accumulating from the obstruction of the surface drainage. On the Mississippi, great waves were created which overwhelmed many boats and washed others high upon the shore, the return current breaking off thousands of trees and carrying them out into the river. High banks caved and were precipitated into the river; sand bars and points of islands gave way; and whole islands disappeared" (M. L. Fuller). Within a year after these first great shocks, hundreds of other shocks were felt, one on January 23, and several on February 7, 1812, having been very severe. Some of the shocks were felt on the Atlantic seaboard. It seems quite certain that these earthquakes were caused by slipping or readjustment of earth blocks along a general line or zone of fracture in the older rocks lying underneath the unconsolidated sediments of the Mississippi River flats.

Chilean earthquakes. Chile has been visited by a number of violent and destructive earthquakes during the last 200 years. Those of 1822 and 1835 both shook hundreds of thousands of square miles of the southern part of South America, and, just after each, long stretches of coastline were found to be elevated several feet. Those of 1868 and 1922 both caused great tsunamis to rush upon the land with destructive violence. In 1906 havoc was wrought in Valparaiso and vicinity, with after-shocks continuing for a long time.

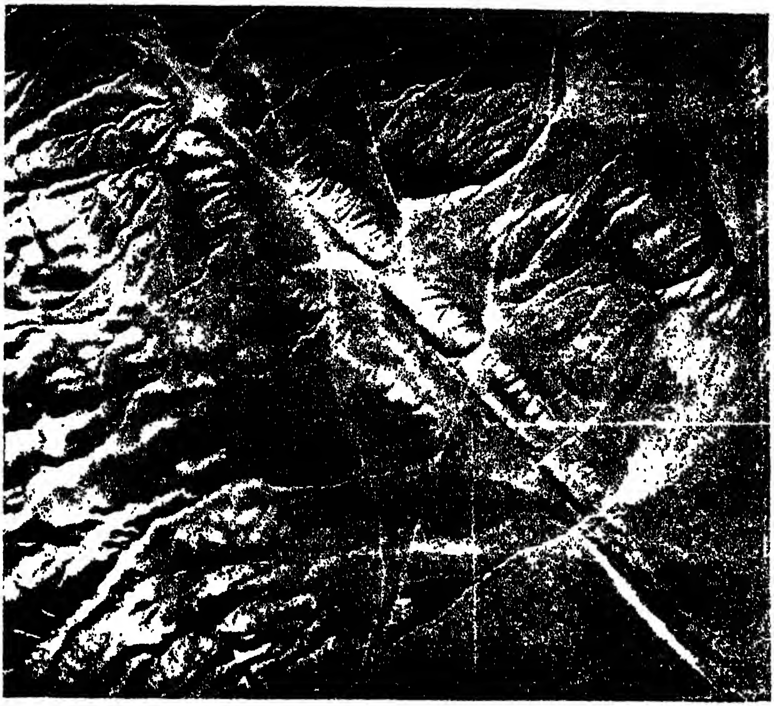


FIG. 57. An aerial view of part of the San Andreas fault trace or rift on the western side of the San Joaquin Valley, California. Note the distinct offset produced by cross-faulting. (Fairchild Aerial Surveys.)

Charleston, South Carolina, in 1886. A violent earthquake shook Charleston, South Carolina, and vicinity in 1886. "A slight tremor which rattled the windows was followed a few seconds later by a roar, as of sub terranean thunder, as the main shock passed beneath the city. Houses swayed to and fro, and their heaving floors overturned furniture and threw persons off their feet as, dizzy and nauseated, they rushed to the doors for safety. In 60 seconds a number of houses were completely wrecked, 14,000

chimneys were toppled over, and in all the city scarcely a building was left without serious injury (Fig. 56). In the vicinity of Charleston, railways were twisted and trains derailed. Fissures opened in the loose superficial deposits, and in places spouted water mingled with sand" (W. H. Norton). It was felt by people in places as far away as eastern Iowa, Boston, Cuba, and the Bermudas. It was caused by a rupture of the old rocks which underlie the loose Coastal Plain strata.

FIG. 58. Road offset 20 feet horizontally along the San Andreas fault at the time of the California earthquake of 1906. Near Point Reyes. The broken line indicates the fault trace. (After G. K. Gilbert, U. S. Geological Survey.)

Japan in 1891. The great Japanese earthquake of 1891 caused the destruction of 20,000 buildings and thousands of people within one minute. It perceptibly shook an area of over 240,000 square miles, but the principal destruction was confined to a thickly settled valley among the mountains in the central part of the island of Hondo. It was caused by a sudden slipping of as much as 10 to 30 feet along a line of earth fracture for 40 miles (Fig. 49). The land on one side of the fracture dropped below that on the other side, leaving a terrace with a steep front as much as 20 feet high. An average of 500 aftershocks a month for five months succeeded the great earthquake.

Assam, India, in 1897. One of the greatest of all recorded earthquakes took place in the Assam region of northeastern India in 1897. Within two and one-half minutes an area nearly as large as California was laid in

ruins, and notable changes in topography took place. The ground was fissured in many places, and through numerous vents great quantities of water and sand issued. At one place a sharply defined terrace 35 feet high was developed. Movements of the earth's crust along one of the lines of fracture followed a winding stream, causing ponding of its water in some places, and waterfalls in others. The land was thrown into waves, and it moved in a remarkable manner. Many landslides occurred.

Southern Alaska in 1899. A series of very violent earthquakes shook the Yakutat Bay region of southern Alaska in 1899 when one part of the coast rose as much as 47 feet (Fig. 33), while another part sank a little below sea level. "Vast quantities of snow and ice were avalanched from the mountains, and, as a result of this abrupt accession of supply to the reservoirs of the glaciers, a wave of advance was started which, during the succeeding years, swept down the glaciers and caused notable change and advance in the glacier ends" (Tarr and Martin). A tsunami destroyed a forest along the coast.

California in 1906. The California earthquake of 1906 ranks as the most violent shock recorded in the United States since the beginning of the twentieth century. The shock lasted about a minute. It caused a property damage, mainly in San Francisco, of several hundred million dollars, but fortunately the loss of life was not great. It was caused by a sudden horizontal movement of one part of the Coast Range Mountains 0 to 22 feet past the other along a line of fracture (fault) for about 250 miles (Fig. 45). Along the fracture, fences, waterpipes, and roads were notably dislocated (Fig. 58), and the ground was torn up. In San Francisco the greatest damage by far was accomplished by fire which started in various damaged buildings and quickly spread.

Fig. 59. House on the main fault wrecked at the time of the California earthquake of 1906. (Photo by R. L. Humphrey, for U. S. Geological Survey.)

Sicily in 1908. In regard to both violence and loss of life, the Messina, Sicily, earthquake of 1908 ranks as one of the greatest in the annals of human history. It has been estimated that between 150,000 and 200,000 people lost their lives in this frightful catastrophe which was caused by the sudden slipping of the earth's crust along a fault fracture.

Japan in 1923. On September 1. 1923, radio messages startled the

world with news of the frightful earthquake disaster which overtook the region including Tokyo and Yokohama in Japan. Earthquake and fire destroyed a large section of the great city of Tokyo, and Yokohama was almost completely ruined. According to various reports, about 150,000 people were killed, and 100,000 were wounded. About 500,000 houses

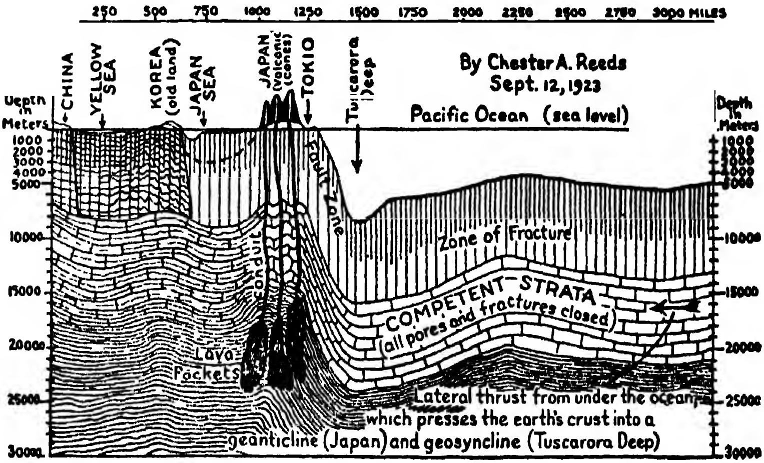


FIG. 60. Diagrammatic east-west cross section of the earth's crust in the latitude of Tokyo, Japan. (After Reeds.)

were destroyed, most of them by fire. The earthquake lasted five minutes. It was caused by a sudden shifting of the earth's crust, said to have been hundreds of feet, along a fault in the bottom of Sagami Bay. Parts of the shores of the bay were elevated 6 feet. According to Shepard much of the change in the configuration of the bottom of the bay was due to submarine sliding seaward of large masses of soft sediment, induced by the shock, rather than to the actual dislocation of the earth's crust.

From the main island of Japan, which rises thousands of feet above sea level, there is a remarkably great and steep descent within a short distance to very deep ocean water (depth about 5 miles). This great, steep slope marks a portion of the earth's crust which is unusually lacking in equilibrium and hence subject to rapid earth-crust movements. Accompanying Fig. 60 gives a general idea of the conditions in this great zone of unusual diastrophic activity.

CHAPTER V

STRUCTURE OF THE EARTH'S CRUST

INTRODUCTION

THE preceding chapter on diastrophism emphasized the nature and evidence of movements of the earth's crust which have taken place in historic and recent times. Attention was called to the fact, however, that throughout the vast span of the geologic past, the crust was always unstable and that the record or proof of such is to be seen in the position and arrangement of the rock layers. Examination of the bedrock reveals many, many, situations in which the layers are no longer in their original, simple positions and shapes, but, instead, are folded, warped, cracked, or displaced into various types of secondary forms, called geologic *structures,* or structural features. That branch of geology which is especially concerned with these features, as well as with the arrangement and relative positions of rock masses, is called *structural* geology. It has become especially significant in recent years because of the known relationships between certain geologic structures and the occurrence of valuable natural resources such as petroleum and ores, and because of the need for understanding these in geologic prospecting and exploration.

Classification of Geologic Structures. In the broadest sense, structural geology embraces a study of all matters pertaining to the arrangement or architecture of the rock materials of the earth. As such it is concerned with the conditions of the deep interior as well as the features of the outermost rock layers. Thus might be included some of the structural "peculiarities" of the outer rocks which are inherited at the outset, such as unusual features of sedimentation like ripple marks, cross-bedding, and concretions. However, the chief concern is for those irregularities, heretofore mentioned, which are wholly or in part the products of diastrophism and which, consequently, are mainly secondary in origin. In this category fall the main types of geologic structures to be considered here, namely, (a) *folds* or flexures, (b) broad warps, (c) joints, (d) faults, and (e) unconformities. Because, in their emplacement, they are more or less related to diastrophism, the igneous rock bodies constitute a special group of structural features to be in-

cluded herein. Before describing these structures it is first appropriate to give consideration to certain factors which are involved in their study and understanding.

Structure Sections. It is usually an important part of the business of the geologist, in reporting on a region, not only to make a geological map depicting the surface distribution of the various rock formations, but also to represent graphically the underground relations of the various formations by means of structure sections (Figs. 61, 62).

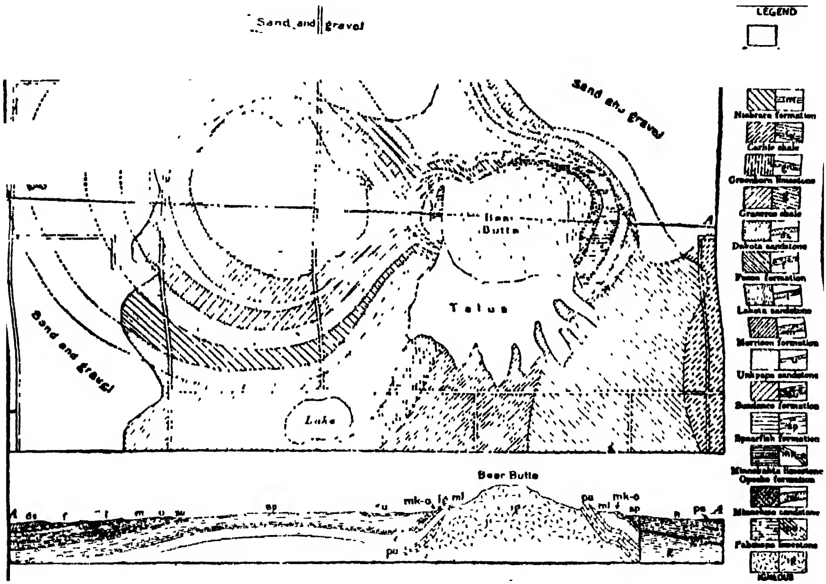


FIG. 61. A geological map and the structure section along the line AA. Bear Butte region, South Dakota. (By C. C. O'Harra, U. S. Geological Survey.)

A *structure section* shows the relations of the rocks of a region as they would appear, from the surface downward, on the face of a vertical slice through a part of the earth's crust along a certain line. Thus, in Fig. 61 the structure section shows the subsurface positions of the various formations along the line AA of the accompanying areal geologic map. A picture of part of the area (Bear Butte) is shown in Fig. 114. In a diagram such as Fig. 62, it is feasible to combine with structure

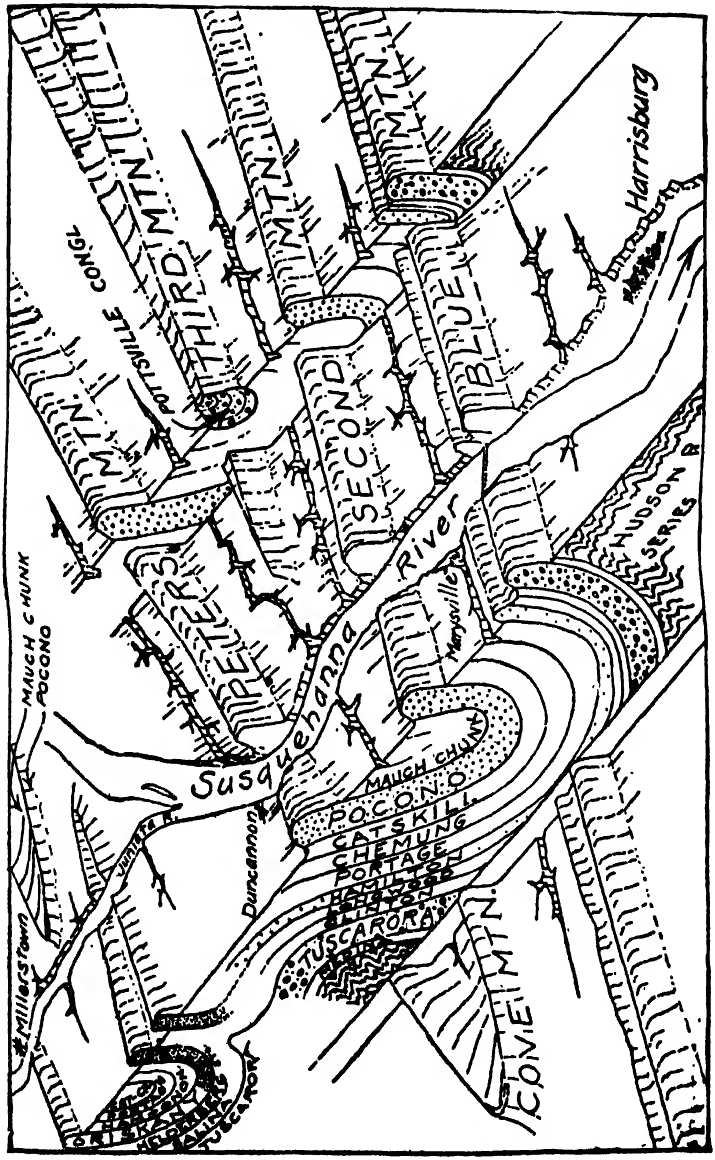


FIG. 62. Diagram illustrating structure sections through the folded, mountainous region west of Harrisburg, Pennsylvania. (Drawn by A. K. Lobeck.)

sections both the surface distribution of formations and the relief of the surface.

Outcrop. Several terms are very commonly employed in dealing with the arrangement of the rocks of the earth's crust. One of these terms is *outcrop,* which means any surface exposure of the bedrock. The term *rock exposure* is sometimes used as a synonym. It so happens that the bedrock formations are, in most regions, largely concealed under cover of soils, loose rock fragments, glacial deposits, vegetation, water, ice, or snow. In high mountains, or in other regions where erosion has proceeded rapidly, outcrops are generally much more numerous and extensive than in regions where sediments have recently been, or are being, deposited. It is very generally true that the surface distribution of rock formations and the underground structures of a region are worked out by a careful study of the outcrops. Where the geologic structure is simple, relatively few outcrops may suffice in order to determine the structure, but where it is very complex many outcrops must be found and carefully studied.

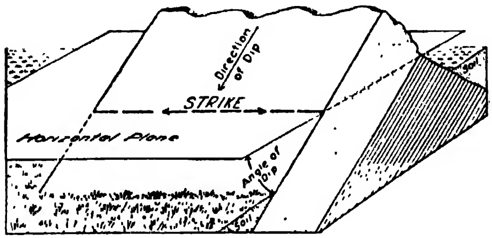


FIG. 63. A block diagram illustrating dip and strike in tilted strata.

Dip and Strike. In many regions, particularly in mountains like the Appalachians, the Rockies, and the Sierra Nevada, rock layers and formations are not only variously tilted or inclined, but also they show marked deviations in trend across country. Two terms are used to designate such variations in tilt and trend—dip and strike. *Dip* is the inclination of a rock layer from a horizontal plane (or level surface). Two elements are involved, namely, *amount of dip* and *direction of dip.* By means of a compass provided with a small pendulum free to move over a graduated half-circle, amount and direction of dip are determined. Examples of note-book records would be dip 20°, S.40° W.; dip 65°, N.10° E., etc. *Strike* is the line of intersection of a dipping or tilted layer with a horizontal plane (or level surface). Or it may be defined as the direction (or trend) of a horizontal line on the surface of an outcropping rock layer. Strike is recorded thus: N.30° W., S.80° E., etc. When the direction of dip is recorded it is not necessary to give

the strike because dip and strike are always at right angles, as clearly shown by Fig. 63. However, in practice, it is generally more easy first to determine direction of strike in the field to be followed by computations of the dip.

Folds

Cause of Folds. A *fold* is a bend in a rock layer or formation. Any rock mass, when subjected to sufficient pressure or stress within the earth's crust, must either bend, fracture, or be mashed. Rock bends are folds, and fractures are joints, fissures, or faults. Bending and mashing, or flattening, are both comprised under the term *rock flowage*, which means a permanent change of form of a rock by pressure but without notable fracture. Why do rocks sometimes bend and at other times break? The earth has, for many millions of years, been a shrinking body. Many stresses, strains, and pressures have been set up in the crustal (outer) portion of the earth as it has been adjusting or accommodating itself to the contracting interior. Because of such forces, the rocks at and near the earth's surface have, in many places, been more or less profoundly fractured and often subjected to sudden movements, while the rocks well within the crustal portion, that is, usually from some thousands of feet to miles down, have in many places been folded or even crumpled. For such reasons the surface and near-surface portions of the crust may be, in a general way, called the *zone of fracture*, whereas deeper portions may be called the *zone of flow*. Rocks (even the hardest) behave like plastic materials when subjected to great pressure well within the zone of flow, and, therefore, they bend or flatten out instead of break, because of the great weight of overlying material. This conclusion has been confirmed by laboratory experiments in which small masses of rocks have been subjected to slow, differential pressure equivalent to that which obtains miles within the earth. Such masses have been made to change shape notably without fracture.

The idea of depth should not, however, be too much emphasized in considering the zones of flow and fracture because there are other conditioning factors. Thus, very soft, plastic rock materials like wet clay at or near the surface will flow readily under pressure, whereas very hard, rigid rocks like granite will usually flow only at depths of miles. Again, a very slowly applied pressure may cause a relatively hard rock to flow or bend much nearer the surface, whereas a quickly applied pressure may cause a relatively soft rock to fracture at a considerable depth.

Kinds of Folds. As already stated, a rock bend is a *fold.* A simple arched-up fold is an *anticline* (Fig. 66). A *syncline* is an inverted anticline, or a downfold (Fig. 70). The flanks of the fold are its

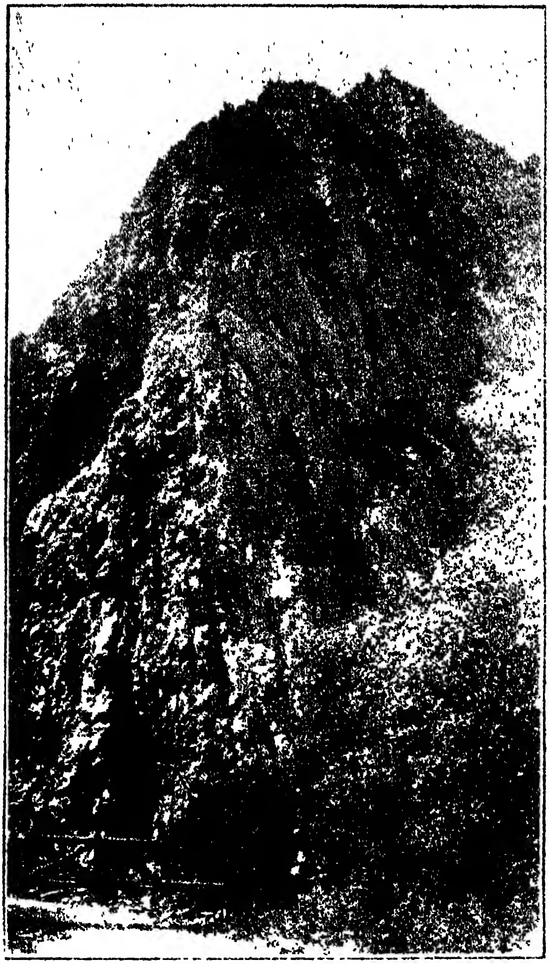


Fig. 64. Steeply inclined and deeply eroded beds of quartzite many millions of years old in the much faulted and folded Wasatch Mountains near Ogden, Utah. (Courtesy of the Ogden Chamber of Commerce.)

limbs, and the crest line (or the trough line) is its *axis.* The inclination of the axis from the horizontal plane is the *pitch* or *plunge.* All these features, as well as dip and strike, are represented by the accom-

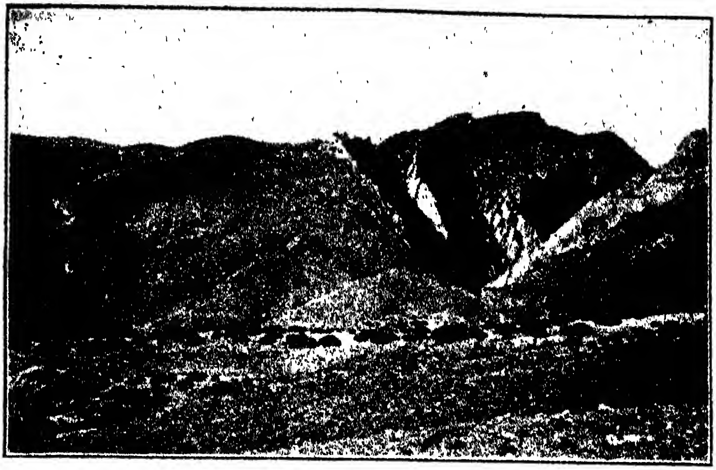


FIG. 65. An anticlinal fold with a strong downward plunge. Bighorn Mountains, Wyoming. (After Darton, U. S. Geological Survey.)

FIG. 66. An anticline in Paleozoic strata near Hancock, Maryland. (After Walcott, U. S. Geological Survey.)

panying diagrams (Figs. 67 and 68). The imaginary plane that bisects a fold is called the *axial plane*. The axis passes along it.

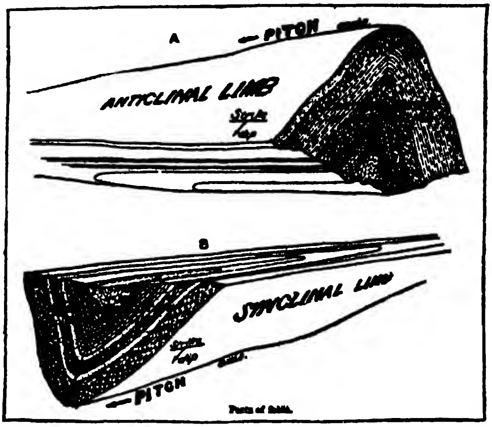


FIG. 67. Diagrams illustrating parts of folds. (After Willis, U S. Geological Survey.)

When a fold has but one limb, that is, when its layers are bent in one direction only, it is called a *monocline*.

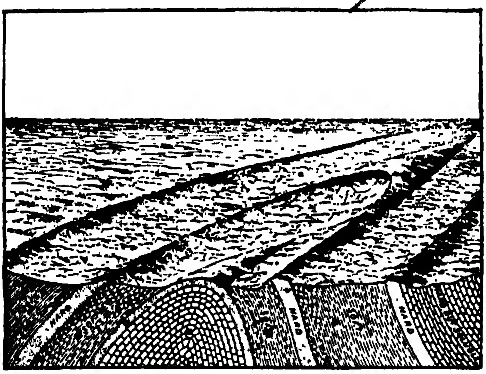


FIG. 68. Perspective view and structure section illustrating a pitching anticline. (After Willis, U. S. Geological Survey.)

In an *isoclinal fold* or a series of such folds, the limbs are parallel or nearly so. Such folds indicate great degrees of pressure (Fig. 71). An *overturned fold* is one in which one limb is partly doubled under the other and the axial plane is inclined (Fig. 72); a *recumbent fold*

FIG. 69. Strata sharply folded into a series of anticlines and synclines. Calico Hills near Daggett, California.

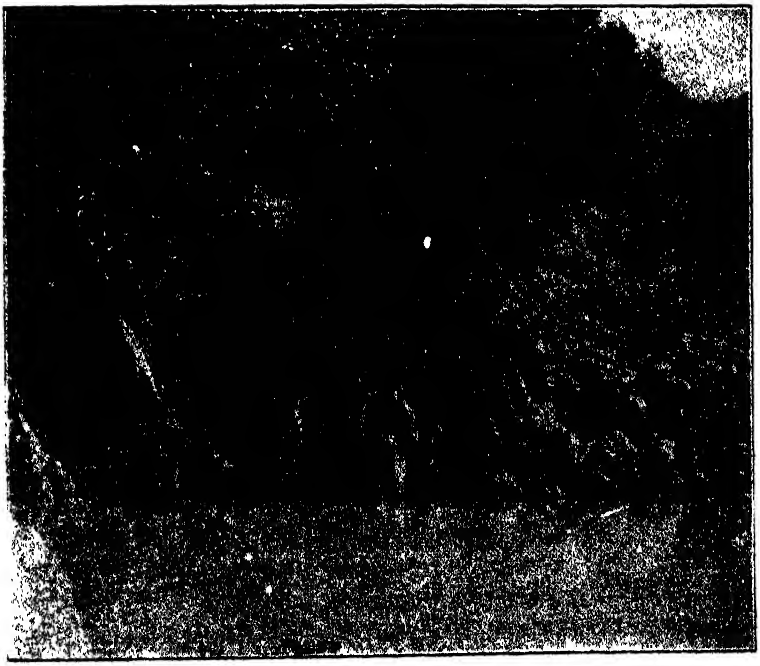


FIG. 70. A syncline near Hancock, Maryland. (After Walcott, U. S. Geological Survey.)

is an extreme case of overturning in which the axial plane lies in a horizontal position.

The terms *anticlinorium* and *synclinorium* may be used to designate, respectively, great, complex, anticlinal and synclinal structures.

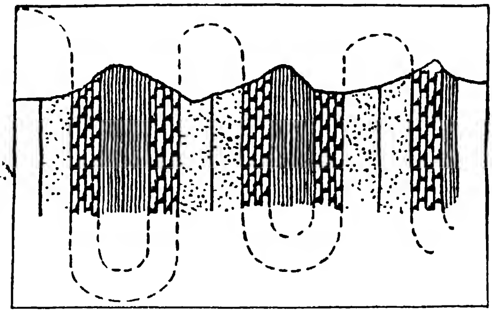


FIG. 71. Diagrammatic structure section showing isoclinal folds. (After Van Hise.)

The term *dome* is sometimes applied to a special type of anticline in which the axis is nearly or quite reduced to zero, that is, the limbs dip downward in all directions from the top of the fold. A synclinal *basin* is an inverted dome.

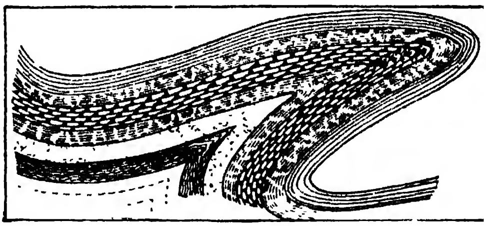


FIG. 72. Diagrammatic section of an overturned fold (After Van Hise.)

Examples of Folds. Folds range in both width and length from almost microscopic to many miles. Most folds, especially the large ones, were developed very slowly, that is, the process may have required thousands or even hundreds of thousands of years. Figs. 73 and 74 illustrate small-scale folds or contortions. The Uinta Range of northern Utah, many miles long and wide, is essentially a great, simple anticline whose limbs dip at very low angles. The Jura Mountains between Switzerland and France contain a series of moderately folded, symmetrical, little

eroded anticlines and synclines (Fig. 335). In parts of the northern Rockies of the United States, the anticlines and synclines are considerably squeezed together and rather deeply eroded. The Appalachian

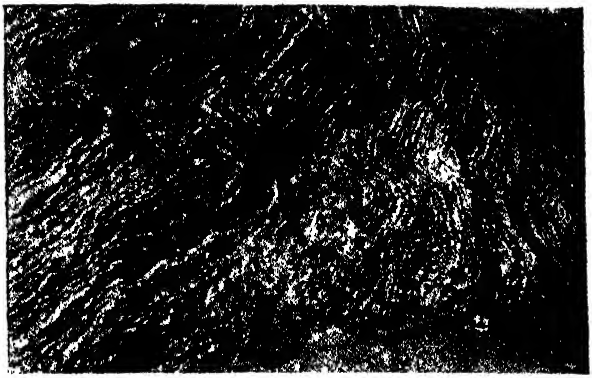


FIG. 73. Contorted beds of shale. Los Angeles, California.

region exhibits almost all known kinds of folds and on almost all scales up to miles across. In Pennsylvania the rocks are less severely folded than they are in the southern Appalachians where they are more closely

FIG. 74. A small, overturned fold, a few feet across, in thin beds of shale. Near Laguna Beach, California.

folded and are more faulted (Fig. 336). The rocks of the Alps have been so severely folded as to give rise to an almost unbelievably complicated structure which is in part suggested by Fig. 338.

FIG. 75. A small double fold. Monterey National Forest, California. (Courtesy of the U. S. Forest Service.)

BROAD WARPS

The earth's crust may be gently *warped* by diastrophic forces which operate vertically, causing some areas to rise while others sink. Such movements are usually uneven, causing the warped surfaces to be irregular. In all cases, however, dips of warped surfaces, or dips of warped strata, are so slight that they are usually very locally unrecognizable. Broad-scale *tilting* is a special case of warping where gentle slopes or dips are very uniform over considerable areas. Warps are usually broad,

often involving distances of 50 to many hundreds of miles, but smaller warps are not uncommon.

Willis, in his book on "Geologic Structures," distinguishes between warping and folding as follows: "In general usage the broad departures of strata from a flat attitude are called folds, as are also more pronounced bends. There is a difference, however, between the two kinds of forms. The broad type . . . is due to vertical movements, to subsidences, or to uplifts. The more pronounced bends, such as commonly occur as alternating troughs and arches, are caused by horizontal compression. When both kinds are designated by the same name a confusion of ideas results and may lead to wrong inference. It is better

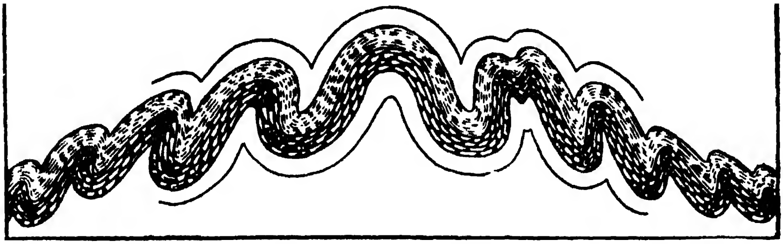


FIG. 76. Diagrammatic section of an anticlinorium. (After Van Hise.)

to call the effects of warping *flexures*, and to restrict the terms folding and folds to the mechanical disturbances caused by compression."

Gentle upbends are called *upwarps*, and gentle downbends are called *downwarps*. Great upwarps, involving dimensions of scores or hundreds of miles, are known as *geanticlines*, and great downwarps of similar size are known as *geosynclines*. The latter are of particular geologic interest because at various times, and at various places, during the earth's history they have been sites of accumulation of very extensive piles of strata reaching total thicknesses of many thousands of feet. Such sediment-filled geosynclinal basins have usually been subjected to tremendous lateral pressure, strongly folded and raised into mountain ranges.

A few examples of large warped regions will now be mentioned. Large parts of northern Europe (e.g., Sweden) have gently risen hundreds of feet while others have subsided in recent geologic time, and it is quite certain that such movements are still more or less active. The region of New York, New England, and southeastern Canada has been upwarped to a maximum of hundreds of feet since the Ice Age, that is, in a period of about 12,000 to 15,000 years. Underlying the

Colorado Plateau of northern Arizona and eastern Utah, a great series of stratified rocks has been regularly warped in many areas. The so-called "San Rafael Swell" in northeastern Utah is a fine example. The strata underlying the Atlantic Coastal Plain of the United States are extensively tilted with very gentle dips, but with comparatively little warping.

Joints

Nature and Occurrence of Joints. Almost all consolidated rock formations at and near the earth's surface are intersected by systems of frac-

Fig. 77. Nearly vertical joints in granite. Near Paradise Valley, South Fork Kings River Canyon, California.

tures or cracks which divide the rocks into blocks of varying size, shape, and spacing. Where no displacement of their adjacent walls has taken place, such cracks are called *joints*. In many places there are at least two sets of joints crossing each other at high angles, and dividing the rock into prismatic blocks of roughly uniform shape and size (Figs. 79 and 82). In many other cases, however, numerous joints traverse rock formations very irregularly. The spacing of joints varies from a fraction of an inch to many yards. Single large joints are hundreds and even thousands of feet long and deep, as may often be seen in steep

canyon walls and great cliffs. It seems quite certain that the whole zone of fracture of the earth is more or less jointed. Joints cannot exist below depths of approximately 15 or 20 miles because, as demonstrated by experiment, the tiniest cracks and crevices in even the hardest rocks cannot remain open under the conditions of tremendous pressure which obtain below such depths. Joints very often stand in approximately vertical positions, but they may lie in any position from vertical to even horizontal, especially when the rocks containing them have been disturbed by folding or tilting. Joint faces often are remarkably smooth

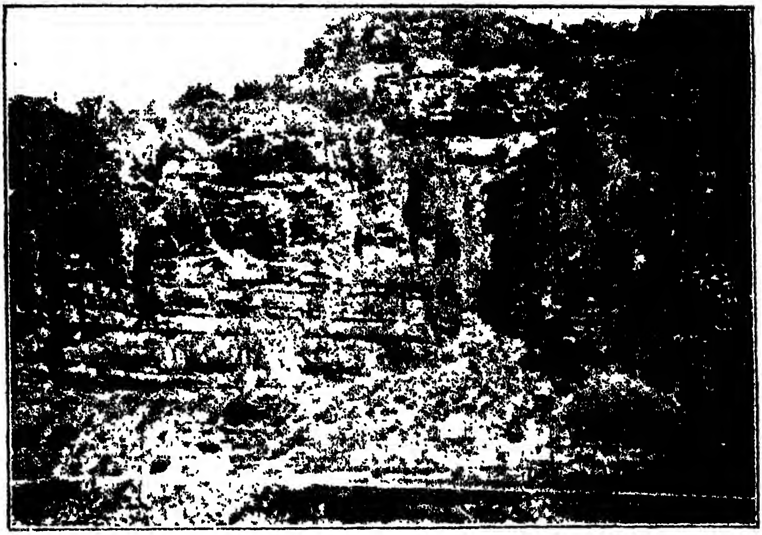


FIG. 78. Vertical joints transecting horizontal sandstone strata. Canyon of the Colorado (Grand) River. Near Glenwood Springs, Colorado.

and straight for long distances, particularly in fine-grained, hard rocks. Joint blocks are usually fitted together tightly, but sometimes there are very perceptible spaces between them, especially if the joints have been acted upon by the weather.

Causes of Joints. Most joints in sedimentary, igneous, and metamorphic rocks are believed to have resulted from stresses within the zone of fracture causing rocks to be so strained that they fracture. Such joints may, on the basis of origin, be classified as *tension* and *compression* (or *shear*) *joints.* When a portion of the zone of fracture is subjected to differential compression or torsional stress, owing to earth contraction,

the rocks tend to fracture in two general sets of joints approximately at right angles to one another very much as can be shown by experiments with glass or ice. The crest portions of folds, where not too deeply buried, are often stretched to the breaking point, resulting systems of joints. The sudden alternation of tension and compression in the zone of fracture during the swift passage of earthquake waves is quite likely a cause of many minor joints. Tension produced by contraction during the drying-out and consolidation of sediments when raised into land also is probably a minor cause of jointing. In igneous rocks, systems of shrinkage joints no doubt often develop when the masses of molten materials contract during the process of cooling and solidification within the earth's crust. Cracks thus formed are often subsequently filled with molten material forced up from greater depths to form what are called "dikes."

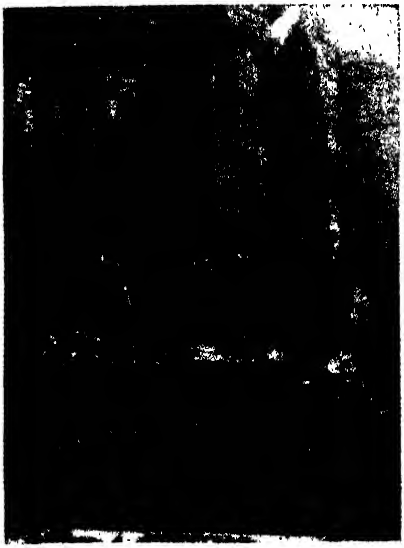


FIG. 79. Great joint columns 1500 feet high in sandstone. Zion Canyon, Utah.

A kind of jointing of special interest is known as *columnar structure.* It develops by cooling and contraction of either lava flows or masses of molten material which have been forced into fissures near the earth's surface, during the process of solidification of the molten material. The effect is a splitting up of the body of rock into a system of close-fitting prisms of varying size and shape but usually hexagonal. They are of all sizes up to several feet in diameter, and 200 or more feet in length. The columns always form at right angles to the cooling surface, so that in lava flows they are vertical, or nearly so, and in dikes they are usually approximately horizontal. In some places this columnar structure is developed on large scales with a wonderful degree of regularity, as for example at the Devil's Postpile in California (Fig. 80); in the Columbia River Canyon; at Devil's Tower in Wyoming; and at the Giant's Causeway in Ireland. The Palisades of the Hudson near New York

FIG. 80. Remarkably developed columnar structure in lava. Devil's Postpile National Monument, California. (Courtesy of the U. S. Forest Service.)

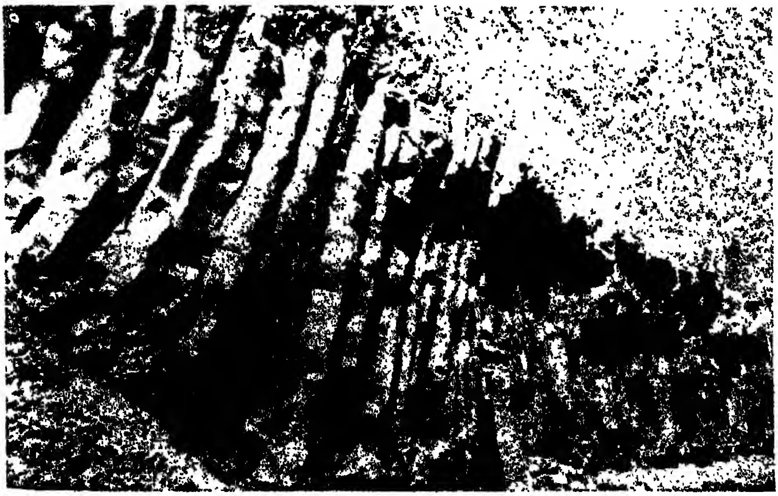


FIG. 81. Crudely developed columnar structure in lava. Yellowstone National Park.

FIG. 82. Highly jointed plutonic rock (syenite). Near North River, New York.

City are great, crudely developed, nearly vertical joint prisms 100 feet or more high.

Sheet jointing is sometimes well developed in massive rocks like granite and other plutonic rocks. The joints usually divide the rock into a series of crude sheets roughly parallel to the surface of the earth. The sheets commonly range in thickness from less than a foot to about four or five feet (Fig. 83). Well below the surface a massive rock body may be in a strained condition. Removal of the overburden by erosion, or even by artificial means, may relieve the strain with resultant splitting loose of successive sheets of the rock from the surface down for 50 or 100 feet or more. Such sheets are known to have broken loose suddenly in granite quarries.

FAULTS

Nature and Occurrence of Faults. A *fault* is a fracture in the earth's crust along the face of which there has been slipping (or displacement) of the rocks (Figs. 84, 85). The amount of such displacement may vary from a fraction of an inch to many thousands of feet. Movement along a fault generally takes place suddenly, commonly involving distances up to 20 to 40 feet, or rarely even more, at one time. In great faults the displacement represents the sum-total of many relatively minor, sudden movements. In some, or possibly many, cases movements may have been so slow as to be imperceptible to the senses. Recent studies suggest that these slow movements may be more common and important than formerly suspected. One interesting case may be mentioned. On the western side of the San Joaquin Valley of California several oil wells were drilled across a fault far below the earth's surface. After some years, due to slow movement along the fault, the well holes

were sheared off to such an extent that the wells had to be abandoned. Sudden slippage along faults is the principal cause of earthquakes as already pointed out.

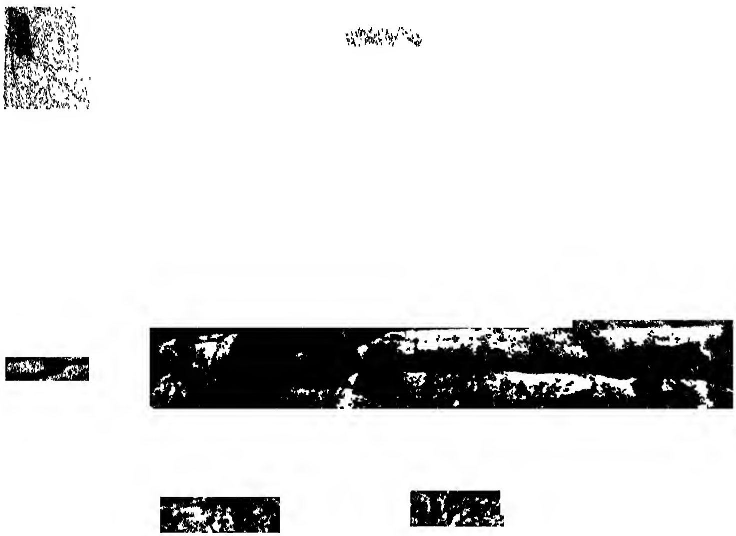


FIG. 83. Sheet jointing in plutonic rock accentuated by weathering. Near Twenty-nine Palms, California.

Components of Faults. Faults are by no means always simple, and, for a reasonable understanding of them, it is necessary to know the principal terms applied, particularly to the amount and direction of movements. The accompanying diagrams should be carefully studied in connection with the definitions below given, and this should be supplemented with laboratory studies of models and maps, and also, if possible, with field observations. The components of faulting may be most readily comprehended by considering faulted strata made up of layers of various kinds, but of course the same general principles apply to faults in any kind of rock.

The *fault surface* is the rock face along which the slipping or dislocation of the rocks occurs. It is better to call it a surface than a plane because it is seldom smooth and straight for any considerable distance.

Slickensides are the smoothed or scratched portions of a fault surface resulting from friction of movement of one earth-block past the other.

• *Fault breccia* is the crushed, broken, and often recemented rock material commonly found along faults, especially the larger ones, due to friction during movement. It often forms crush-zones in brittle

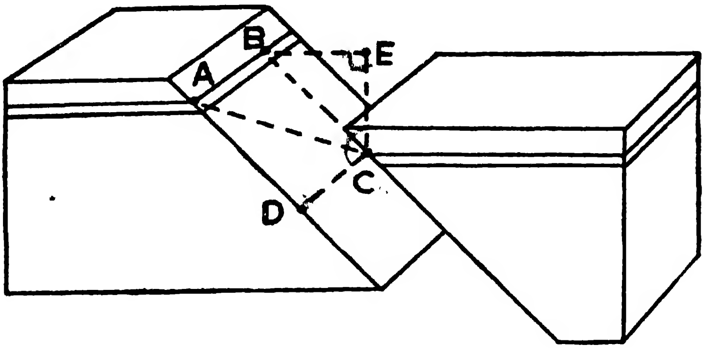


FIG. 84. Block diagram to illustrate various components of a fault, unaffected by erosion, in horizontal strata. ABCD = part of the fault surface; angle CBE = dip; line AB or DC represents strike or trend of the fault; AC = real displacement or slip; BC or AD = dip slip; AB or DC = strike slip; BE = heave; CE = throw; block on left is upthrow side; block on right is downthrow side.

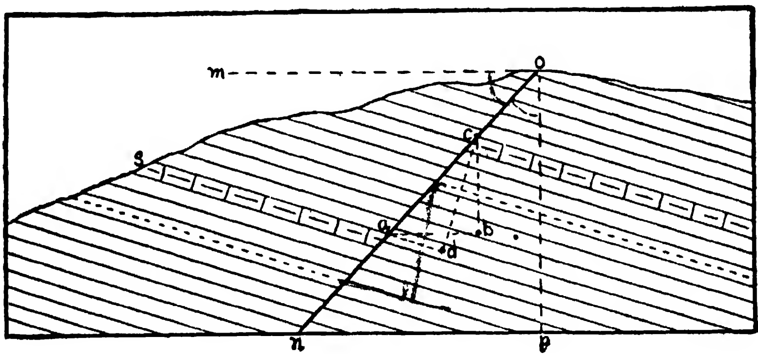


FIG. 85. Structure section of an eroded normal fault in tilted strata. *mon* = dip; *nop* = hade; *ac* = slip; *cb* = throw; *cd* = stratigraphic throw; *ab* = heave.

rock (Fig. 86). In many cases, however, faults are relatively clean, sharp breaks, without breccia separating the bedrock surfaces.

• *Drag* is the term applied to the local bending of strata upward or downward adjacent to the fault surface according to direction of move-

ment of the earth blocks. It is due to friction of the slipping mass along the fault. Drag is by no means always present.

Where one block of earth has moved down relative to the other along a fault, it is called the *downthrow side,* and the other is called the *upthrow side.*

The rock face immediately overlying a fault is called the *hanging wall,* and that immediately under it is called the *footwall.*

Fig. 86. Detail view of fault breccia produced by crushing of hard, brittle rock. San Gabriel Mountains, California.

The *dip* of a fault, like that of a rock layer, is the inclination of its surface to a horizontal plane (or level surface).

The *strike* of a fault is its intersection with a horizontal plane.

The *hade* is the inclination of the fault surface to a vertical plane. It is always the complement of the dip.

Slip is the distance a rock layer has moved on a fault surface. It represents the total displacement along the fault.

Dip slip is the amount of slip in the direction of the dip.

Offset or *strike slip* is the amount of horizontal displacement parallel to the strike of the fault.

Throw is the vertical displacement of the fractured ends of a rock layer.

Heave is the horizontal displacement of the fractured ends of a rock layer as measured at right angles to the strike of the fault.

Stratigraphic throw is the thickness of the rock layers lying between the faulted ends of a given layer, as measured at right angles to the layers.

Principal Types of Faults. There are various types of faults resulting from the complex variety of factors which cause them. In the complete analysis the classification of faults becomes quite elaborate, involving several bases. A satisfactory and simple classification, for purposes here, is based upon the apparent movement along the fault surfaces. Accordingly, two fundamental types are most common, namely, (a) *normal* faults and (b) *reverse* faults.

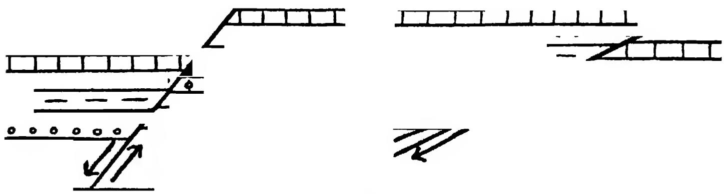


FIG. 87. Structure sections of a simple normal fault (*on left*) and of a simple thrust fault (*on right*), as they would appear if unaffected by erosion.

A *normal fault* is one in which the hanging wall has slipped down an inclined fault surface relative to the footwall. In this case the fault surface dips toward the downthrow side. Fig. 87 (left) illustrates a cross section of a simple case of normal faulting in horizontal strata as it would appear unaffected by erosion on top. Fig. 85 shows a normal fault cutting across tilted beds which have been eroded. As shown by the Figures, normal faulting involves a local extension of the earth's crust because the rocks have been separated, or pulled apart, horizontally by an amount equivalent to the heave. Viewed genetically, this type of fault is frequently referred to as a *gravity* fault and is sometimes called a *tension* fault.

A *reverse fault* is one in which the hanging wall has slipped (has been thrust) up an inclined fault surface relative to the footwall. In this case the fault surface dips toward the upthrow side. The dip of the fault surface may be at a high angle or a low angle. Where the angle is relatively low (less than 45°) evidence may point to consider-

able overriding or overthrusting of the hanging wall, as a result of great lateral pressure. From this genetic viewpoint, such faults are usually called *thrust* faults. In great thrust faults the dip is often less than 20°. Fig. 87 (right) represents a simple case of reverse faulting in horizontal strata as it would appear unaffected by erosion on top. Fig. 98 represents a more complex case where a steeply dipping reverse,

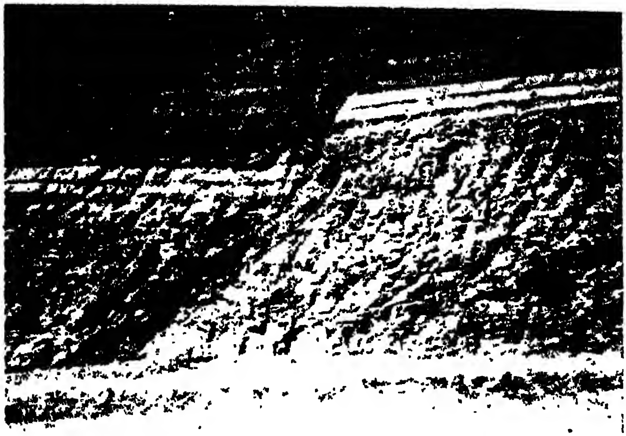


FIG. 88. A small normal fault in shale beds. The displacement is six feet. Downthrow side is on the left. Los Angeles, California.

FIG. 89. Dikes of white granite in dark gray diorite displaced by a normal fault. Near Pasadena, California.

or thrust, fault has been deeply eroded. As shown by these illustrations. the crust of the earth is relatively shortened by reverse faulting since

FIG. 90. A reverse fault showing a displacement of several feet. The rock is sandstone with some thin beds of shale (dark bands). The right-hand side has been thrust upward and to the left. Near Whittier, California. (Photo by M. R. Huberty.)

certain blocks of rock are partly shoved or thrust over others. For this reason, the name *compression* fault is sometimes applied.

Since the relative movements of all faults do not satisfy the condi-

tions of the two fundamental types considered above; additional ones will be described briefly.

Where there has been upward or downward movement on either side of a vertical fault-surface, it is a *vertical fault*. The fault surface dips 90°, or nearly so, and there is neither hanging wall nor footwall.

Where the movement has been wholly horizontal, or nearly so, on either side of a fault surface (either inclined or vertical), it is a *horizontal fault.* A sudden displacement of this kind up to 22 feet along a fault caused the California earthquake of 1906, but such fault movements are relatively rare.

FIG. 91. A nearly vertical fault separating granite (on the right) from contorted beds of quartzite and marble. Southern Inyo Mountains, California.

Where one portion of an earth block moves upward, and another portion of the same block moves downward, on an axis at right angles to the fault, it is a *pivotal fault.* The block on one side of the fault works as though on a pivot with reference to that on the other side. The effect is for the earth block on one side of the fault to be in part upthrown and in part downthrown.

Special Fault Relationships. Two or more parallel, or approximately parallel, faults may be close together. In such cases the total amount

of crustal displacement is distributed instead of being concentrated along a single fault surface. This kind of fault is often referred to as a *compound* or *distributive* fault. Many times two or more normal (gravity) faults may be parallel to one another so that the downthrow is systematically on the same side of all of them, thus producing a step-like pattern of rock displacement. These faults are commonly spoken of as *step* faults (Figs. 92, 95B). A *graben* is a block of the earth's crust, bounded on opposite sides by normal faults, which has been sunk relative to the surrounding blocks (Figs. 95D, 97), whereas a *horst* is

FIG. 92. A normal fault in bedded marl, showing a displacement of several feet, with small step faults on each side of it. Near Lomita, California.

a block of the earth's crust which has been elevated relative to the blocks on either side (Fig. 95C). Grabens and horsts are generally long in comparison with their widths. Frequently they are of such magnitude as to cause conspicuous topographic forms (a discussion of which will follow later).

Causes of Faulting. The zones of flow and fracture in the earth's crust have already been described briefly in the discussion of the cause of folding. Rocks bend (or flow) when subjected to sufficient stress or pressure in the zone of flow, whereas the same forces cause them to break in the zone of fracture. Every fault must, therefore, die out downward because the fracture passes into material which yields without breaking. The forces which cause faulting are compression, torsional stress, or tension (stretching) in the outer crustal portion of the earth. Any such

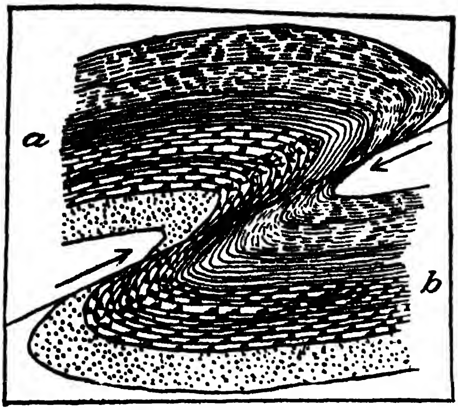


FIG. 93. A sharp fold passing into a thrust fault. (After Van Hise, U. S. Geological Survey.)

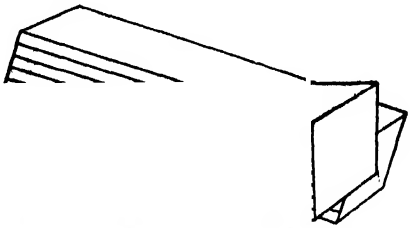


FIG. 94. Block diagram showing the principle of the pivotal or rotational fault.

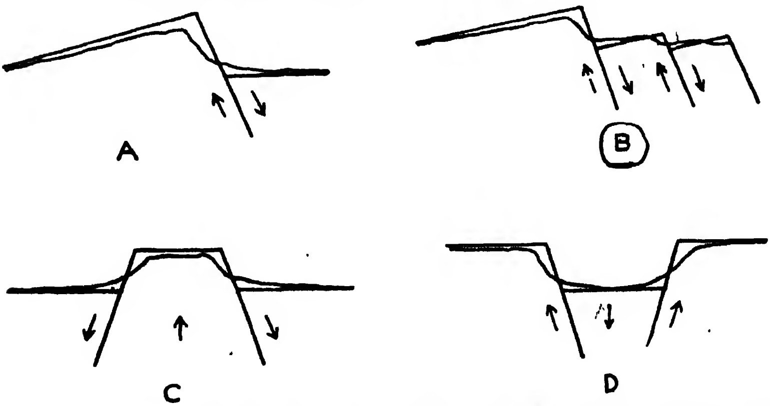


FIG. 95. Diagrammatic structure sections to illustrate various kinds of fault blocks. Straight lines show blocks as if unaffected by erosion and deposition. Other lines are profiles after erosion of high portions and partial covering of low portions with sediment. A, a tilted block; B, step-fault blocks; C, a horst; and D, a graben. Arrows indicate directions of movement.

force may be exerted upon a rock mass until its breaking point is reached, whereupon a fault results, usually accompanied by sudden movement. This movement relieves the force for a time, but the force may increase slowly again to cause renewed movement along the old fracture, and so on repeatedly.

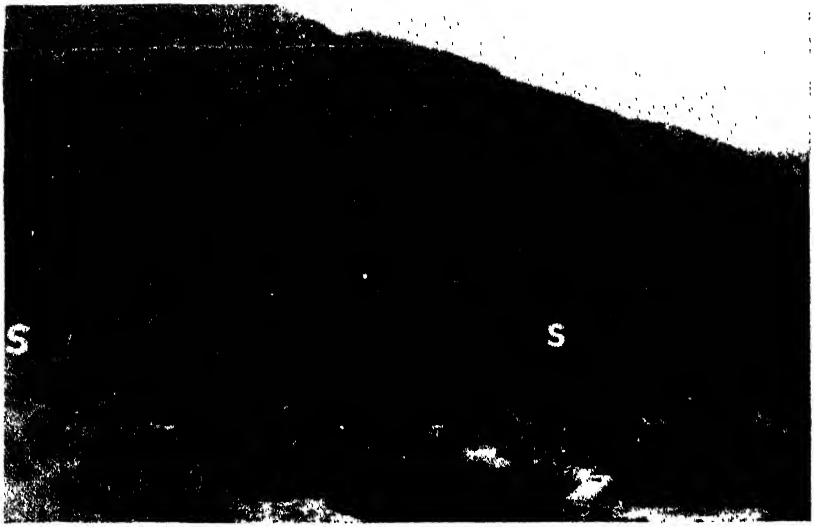


FIG. 96. A normal fault scarp nearly 2000 feet high considerably modified by erosion. The small, slightly eroded scarp SS, near the base of the mountain, was produced by the latest, important, renewed movement along the fault. Deep Spring Valley, California.

Great normal and vertical faults are generally associated with, and seem to result from, upward and downward movements of relatively large sections of the earth's crust, during the movements of which fault

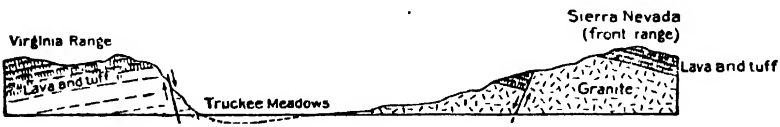


FIG. 97. Structure section of a normal fault-block (graben) which has sunk about 3000 feet in western Nevada. (After U. S. Geological Survey.)

fractures develop. Great thrust faults are very often directly associated with compressive forces which cause zones of the earth's crust to crum-

ple and become elevated into mountain ranges, the thrusts usually developing fairly well up in such masses.

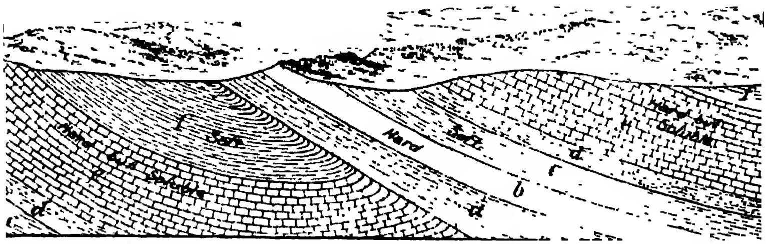


FIG. 98. Structure section and perspective view of a steeply dipping thrust fault. Strata on the right have been shoved leftward (See Fig. 99). (After Willis, U. S. Geological Survey.)

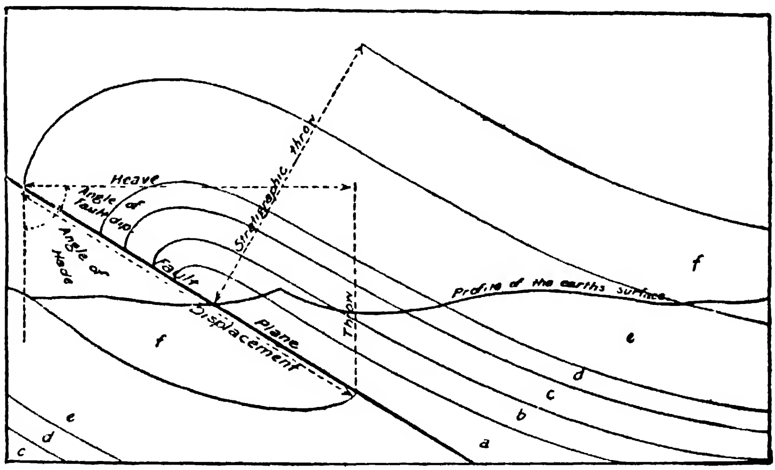


FIG. 99. Diagrammatic section of the thrust fault shown in Fig. 98. (After Willis, U. S. Geological Survey.)

The *ultimate cause* of faulting, as for that of folding, is a more difficult and uncertain matter, but by many geologists, including the author, it is believed to be intimately bound up with a shrinking earth interior, in the outer or crustal portion of which tremendous stresses and strains are, and for countless ages have been, set up.

Topographic Influences of Faulting. Many relief features of the earth, both great and small, are direct or indirect results of faulting. If a considerable fault movement of any kind, except horizontal, should suddenly take place, a cliff or steep slope, called a *fault scarp,* would develop on one side of the fault (Fig. 49). Sudden movement of this kind is, as we have already learned, the most common cause of earthquakes. A single movement rarely produces a scarp more than ten or twenty feet high, but many repeated movements along the same fault

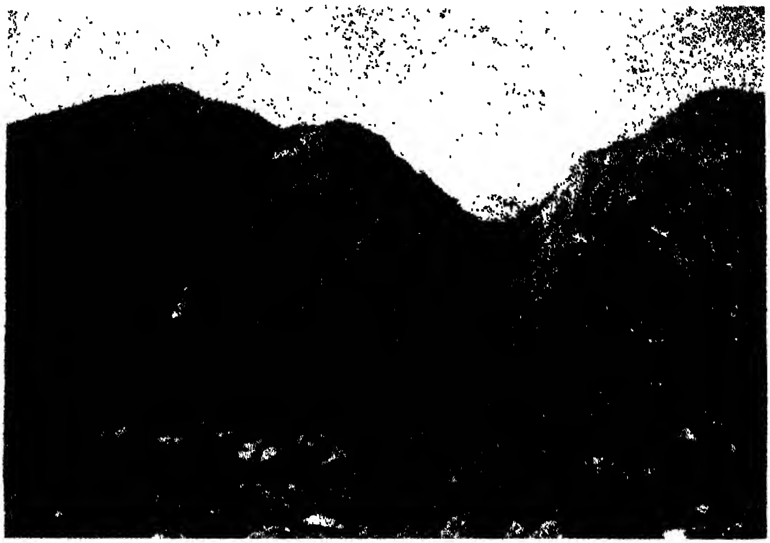


FIG. 100. A fault scarp 1,000 feet high notched with canyons. Base of Inyo Mountains in Saline Valley, California.

may develop a scarp thousands of feet high, as along the eastern face of the southern Sierra Nevada. Even while such a scarp is developing through many thousands of years of time, the processes of weathering and erosion set to work to cut it down and diminish its steepness. Such eroded fault scarps are common in many parts of the world (Figs. 96, 100, 101).

In the course of time both sides of a fault, including the scarp, may be brought to the same level by erosion. If, then, the rock on the upthrow side is weaker than that on the downthrow side, and the whole region should be elevated, erosion would be renewed and the weaker rock would be worn down faster, causing a scarp to develop on the

original downthrow side. Or, if the weaker rock should lie on the downthrow side, the scarp would again develop on the upthrow side. Scarps thus formed by erosion along faults are called *fault-line scarps* by Davis to distinguish them from true fault scarps. Fault-line scarps are common in the eastern Adirondack and Mohawk Valley regions of New York State.

Thrust faulting also often causes great and small changes in topography. Thus the whole eastern face of the Rocky Mountains in Glacier National Park, Montana, is the front of a vast earth block several thousand feet high which has been shoved upon the Great Plains from the west. This fault scarp has been considerably cut into and indented by the action of rivers and glaciers (Fig. 103).

Thrust fault scarps, like normal fault scarps, may be, in the course of time, cut away and fault-line scarps developed instead.

In some regions, like parts of southeastern Oregon, Utah, and Arizona, steplike mountains and ridges, often in series, result from the tilting of fault blocks (Figs. 95B and 345).

A good many valleys of considerable size are the direct results of the sinking of earth blocks (*graben*) between faults. Examples are Death Valley, California; Truckee Meadows, California (Fig. 97); Jordan Valley, Palestine; and the Great Rift Valley of eastern Africa, hundreds of miles long, with its Lakes Albert and Tanganyika.

Many valleys and systems of valleys have been carved out by streams along faults because the work of erosion is, as a rule, more easily accomplished along such lines of weakness in the rocks. A good example of such a drainage system is the southeastern Adirondack region of New York.

Some Examples of Large Faults. The steep eastern face of the Sierra Nevada Range of California is bounded for hundreds of miles by a great, steeply inclined normal fault, the displacement along which has amounted to from one to three miles. In the Owens Valley region, the moderately eroded fault scarp rises very boldly to a height of over two miles (Fig. 203).

Displacement along the great normal fault, fully 200 miles long, at the western base of the Wasatch Mountains of Utah has amounted to many thousands of feet, with 3000 to 5000 feet of it still represented in the face of the steep, moderately eroded fault scarp (Fig. 344).

In southwestern Utah and northwestern Arizona, the great Hurricane fault marks the western boundary of the Colorado Plateau for

about 200 miles. It is a steep normal fault with a displacement of 1500 feet to 6000 feet or more. The downthrow side is on the west. A high, moderately eroded fault scarp marks it throughout (Fig. 101).

FIG. 101. Part of the scarp of the great Hurricane fault. The scarp, 2000 feet high, has been moderately affected by erosion as shown by Fig. 102. About 30 miles south of Cedar City, Utah.

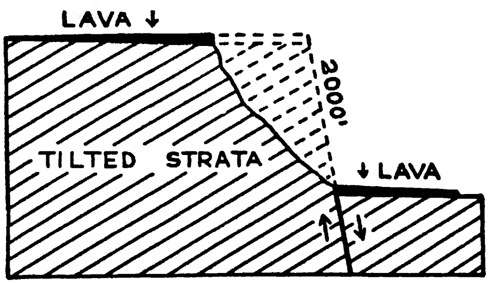


FIG. 102. Structure section through the Hurricane fault scarp pictured as Fig. 101. The lava fields were continuous before the dislocation. The broken lines show the amount of erosion.

Most of the numerous nearly north-south mountain ranges of the great desert region between the Sierra Nevada and Wasatch Mountains are partly eroded, tilted, normal fault blocks.

In the eastern United States, the Adirondack Mountain and Mohawk Valley regions of New York State are cut to pieces by hundreds of vertical or nearly vertical faults, displacements along which are commonly from a few hundred to at least 2000 feet.

Some of the great thrust faults are even more impressive. Thus, several thrusts in the eastern Appalachian region are each 100 miles or more long, with displacements on nearly horizontal or undulating fault surfaces three to five miles or more.

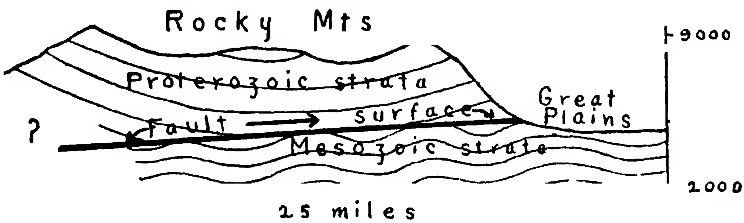


FIG. 103. Diagrammatic structure section of the great thrust fault in Glacier National Park, Montana.

The tremendous thrust fault, so clearly traceable for several hundred miles along the eastern base of the Rocky Mountains of the northern United States and Alberta, Canada, represents the bodily shifting of a large section of the mountain mass eastward over a low-angle fault surface for at least ten to fifteen miles (Fig. 103).

In Scotland, Scandinavia, and the Alps, there are also thrust faults of tremendous magnitude.

UNCONFORMITIES

When strata are deposited in uninterrupted succession, layer upon layer, they are said to be *conformable.* Often, however, there is a break or interruption in the succession of strata which is usually indicated by the fact that one set of strata rests upon the eroded surface of another set. In many cases strata rest upon the eroded surfaces of igneous or metamorphic rocks. The interruption may much more rarely be due simply to lack of deposition of sediments for a time, with no accompanying erosion. Sets of rocks whose regular succession is thus interrupted are said to be ***unconformable,*** and the structure is called an ***unconformity.***

A mass of stratified sediments may be raised out of water and tilted, folded, or left practically horizontal, and then eroded. Submergence would allow new sediments to deposit unconformably upon the eroded surface. Renewed uplift and partial erosion may lay bare such an

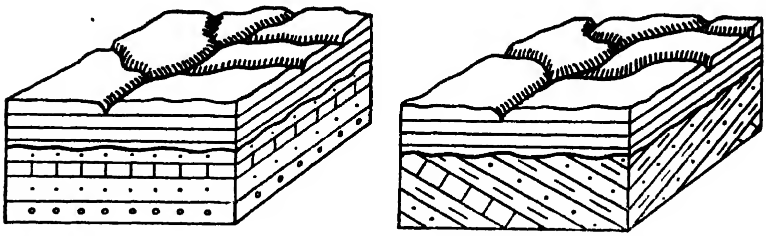


FIG. 104. Block diagrams to illustrate a disconformity (on the left) and a nonconformity (on the right). The heavy, irregular line marks the erosional surface or unconformity.

FIG. 105. A conspicuous unconformity. Horizontal sands and gravels of Quaternary age resting upon tilted and eroded Tertiary shales. Near Port San Luis, California. (After G. W. Stose, U. S. Geological Survey.)

unconformity as illustrated by Fig. 105, in which the underlying strata have been highly tilted and eroded, and by Fig. 342, in which the underlying strata have been notably folded and eroded.

If the upper series of beds rests upon the eroded surface of tilted, or folded, strata or of non-stratified rocks (igneous or metamorphic), there

is a very obvious unconformity called a *nonconformity*. If, however, two sets of beds separated by an erosion surface have their stratification surfaces practically parallel, there is a more or less deceptive unconformity called a *disconformity* (Fig. 104).

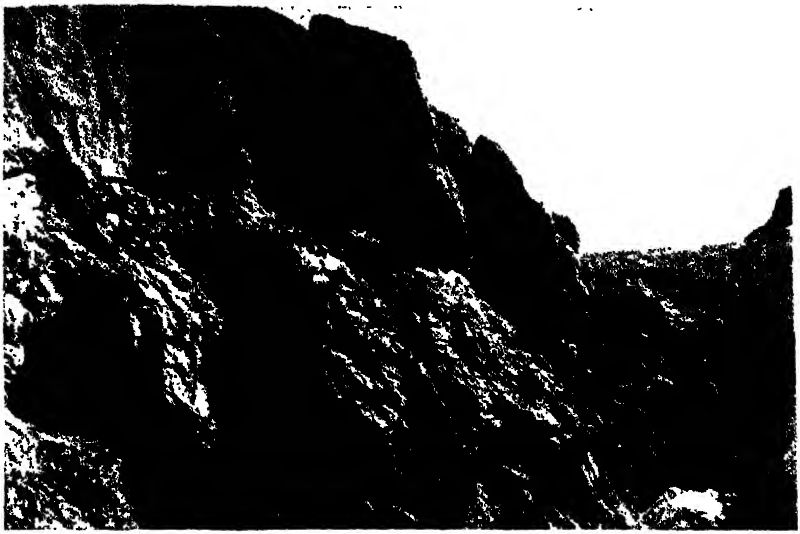


FIG. 106. Horizontal beds of sandstone lying unconformably upon steeply dipping, granite-injected schist. The schist is many hundreds of millions of years older than the sandstone. Painted Canyon in Mecca Hills, southern California.

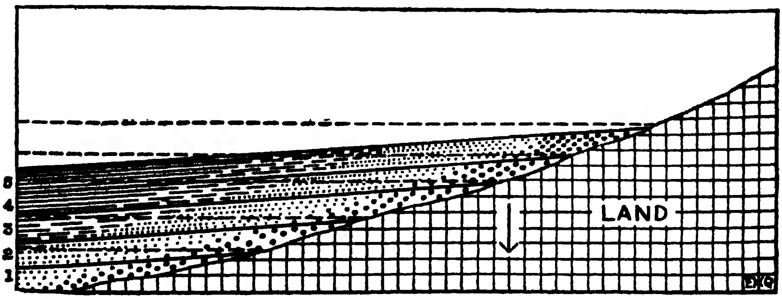


FIG. 107. Diagrammatic section showing the principle of overlap. Horizontal broken lines represent different stages of sea level relative to the land. As the sea encroached toward the right upon the subsiding land, deposition of the sediments 1, 2, 3, 4, 5, extended farther and farther to the right, later-formed beds thus overlapping earlier-formed beds.

A special phase of unconformity is known as *overlap* in which the younger (overlying) strata extend farther, that is, they cover a wider area, than the older (underlying) strata and so overlap the latter. Overlap will develop when strata accumulate upon a sloping area which is gradually subsiding under water. As the sea encroached upon the subsiding land shown by Fig. 107, deposition of the beds of sediments, 1, 2, 3, 4, 5, extended farther and farther to the right, later-formed beds thus overlapping earlier-formed ones.

Emplacement and Structure of Igneous Rocks

Introduction. In even an elementary discussion of the arrangement (structure) of the rocks of the earth's crust, the modes of occurrence of

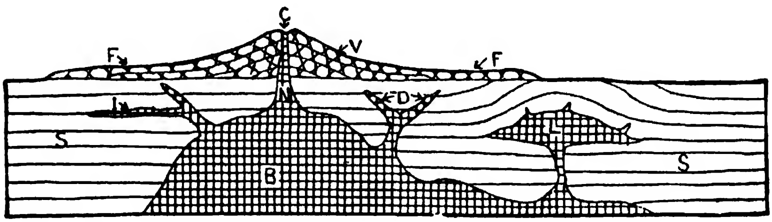


Fig. 108. Diagrammatic structure section illustrating modes of occurrence of igneous rocks. *S* = strata; *B* = batholith of plutonic rock; *L* = laccolith; *D* = dikes; *I* = intrusive sheet or sill; *V* = volcano; *N* = neck of volcano; *F* = lava flows; *C* = crater.

the igneous rocks should be considered because they involve important structural features of the earth's crust (Fig. 108). Studies of igneous rocks in many parts of the world have shown that plutonic (intrusive) rocks are of far greater volume than volcanic (extrusive) rocks. Plutonic rocks become exposed at the surface only as a result of erosion of the rocks which formerly covered them. Volcanic rocks are, no doubt, generally connected with deep-seated plutonic masses through intermediate rocks, reservoirs of molten masses of the latter having always, or nearly always, been the sources of the volcanic materials.

Intrusive Forms of Igneous Rocks. *Dikes.* A *dike* is a mass of igneous rock which, in a molten condition, was forced into a fissure within the earth's crust and there consolidated. Dikes vary in width from less than an inch to several hundred feet, and in length commonly from a few feet to 10 or 20 miles. One in England is about 100 miles long. Dikes are very abundant, a few among many regions being along the

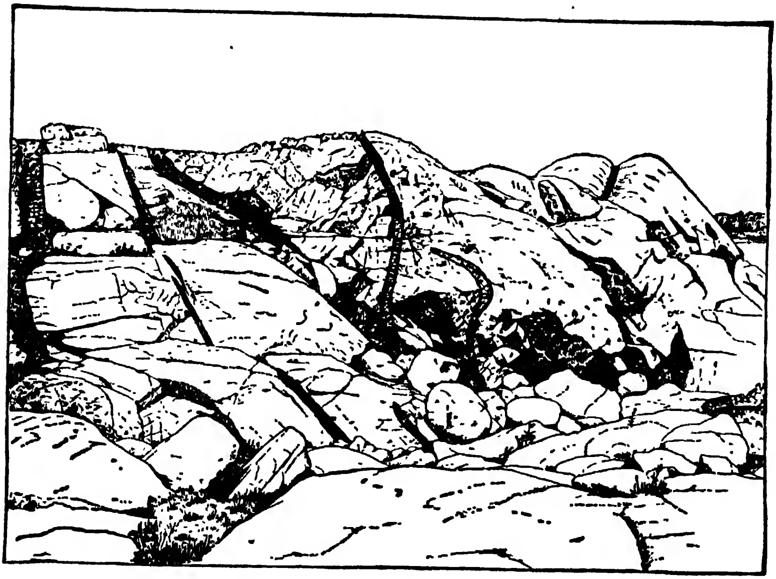


FIG. 109. Basaltic dikes in granite. Cape Ann, Massachusetts. (After N. S. Shaler, U. S. Geological Survey.)

FIG. 110. A small sill of granite parallel to the nearly vertical foliation of schist. A branch of it forms a cross-cutting dike on the left. Seven miles northwest of Palm Springs, California.

coast of Maine; Cape Ann in Massachusetts (Fig. 109); and at Spanish Peaks in Colorado (Fig. 113). Dikes may be partially glassy, fine-grained crystalline or coarse-grained crystalline, depending on the size, rate of cooling, etc., of the injected molten masses. They are sometimes

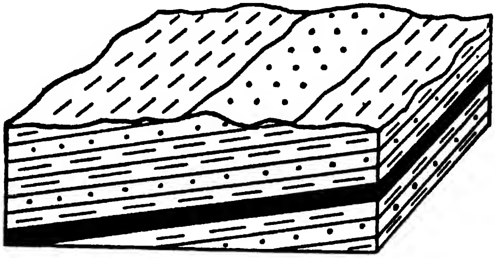


FIG. 111. A block diagram showing a sill (black band) lying between beds of sedimentary rocks.

FIG. 112. Intrusive sheets of granite (white) following the foliation structure of schist (dark). Two miles east of Mountain Spring, Imperial County, California.

arranged in roughly parallel groups, but often they form irregular, branching networks cutting through rocks of any kind.

Sills. Where a mass of molten material (magma) has been forced, in the form of a sheet, between beds of strata, or along well-developed foliation surfaces of metamorphic rocks, it cools to form an *intrusive*

sheet or *sill.* A sill is, therefore, a special kind of dike. Sills may lie in horizontal or inclined positions (Fig. 111). They vary in thickness from less than a foot to hundreds of feet, and they commonly extend laterally from a few acres to many square miles. An excellent example is the sill in the Hudson Valley near New York City, the outcrop of which is called the Palisades of the Hudson, in part forming a bold cliff several hundred feet high facing the river for 30 miles.

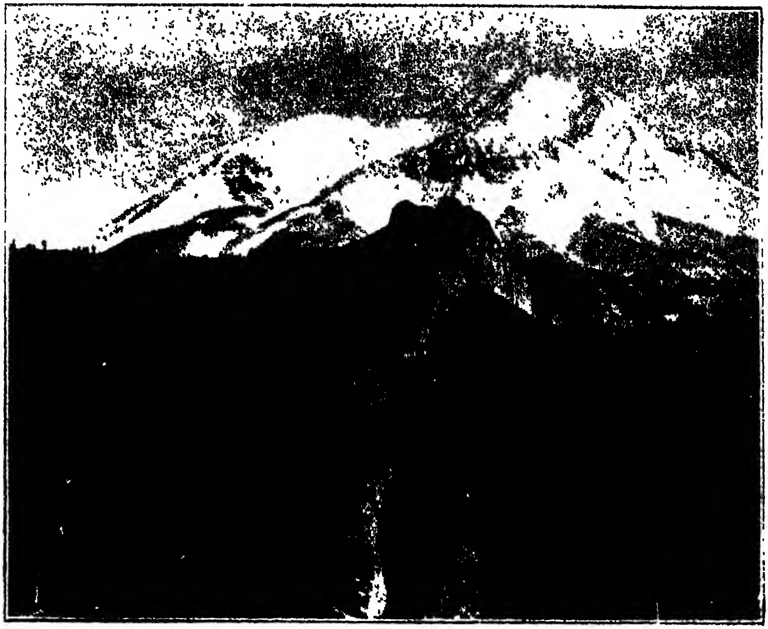


FIG. 113. Vertical dikes cutting strata and standing out in bold relief as a result of erosion. W. Spanish Peak, Colorado. (After G. W. Stose, U. S. Geological Survey.)

Laccoliths. A *laccolith* is a dome-shaped mass of igneous rock which, in molten condition, has been forced between strata, causing the overlying rock layers to be domed or arched up. It has a more or less flat floor. The magma rising into the earth's crust through a relatively small opening becomes such a stiff fluid (or so viscous) that it can no longer penetrate the strata, so it spreads between them and arches up the overlying beds (Fig. 108). Laccoliths commonly range in thickness from a few hundred feet to a mile in the middle, and in diameter from hundreds of yards to several miles. The Henry Mountains of southern

Utah consist of a series of laccoliths showing all stages of removal of the overlying, arched-up strata. Various others occur in Utah, Colorado, Montana, and South Dakota. Bear Butte in South Dakota (Fig. 114) is a very fine example of a large laccolith whose cap rock has been almost completely removed, leaving only the upturned edges of the strata as a ring around its base.

Volcanic necks. A *volcanic neck* is the hardened lava which fills the feeding channel (or conduit) of a volcano. It is roughly cylindrical in

Fig. 114. A laccolith unroofed by erosion. The upturned strata around its base formerly extended completely over the igneous body. Fig. 61 shows it as represented by a geologic map and a structure section. Bear Butte, South Dakota. (After N. H. Darton, U. S. Geological Survey.)

shape, and it commonly varies in diameter from a few hundred feet to a mile or more. Long continued erosion may finally cut away most of the relatively looser material of the volcano, leaving much of the core or neck of the mountain standing out in bold relief. There are excellent examples in New Mexico (Fig. 133) and Arizona and in parts of Great Britain and France.

Stocks or bosses. The term *stock* (or *boss*) is applied to a fairly large body of plutonic rock, with crudely circular or oval ground plan, which, in molten condition, was forced into the earth's crust by cutting across the enclosing rock. Stocks usually increase in diameter downward. They vary in diameter from hundreds of feet to a number of miles. They are very common in New England and in the Piedmont Plateau of the eastern United States.

Batholiths. These are also called *bathyliths* (Fig. 108). In all important respects, except size, they are like stocks. They extend over areas of hundreds to many thousands of square miles, as for example in the southern Sierra Nevada Range; parts of the Rockies; eastern Canada; the Adirondack Mountains; New England; and the Piedmont Plateau. Stocks and batholiths, being true plutonic rocks, are, of course, exposed at the surface only as a result of profound erosion of the overlying rocks. Granite, syenite, and diorite are very commonly the rocks of stocks and batholiths.

Extrusive Forms of Igneous Rocks. *Lavas.* Streams and sheets of molten materials (lavas) may pour out of volcanoes or fissures in the earth and cool to be successively covered by later flows. In such a manner a lava field may be built up to a thickness of hundreds or even thousands of feet and to an extent of many square miles.

Fragmental materials. Through successive explosive eruptions of volcanoes, fragments of lava may be ejected in great quantities and scattered near and far. Thick and extensive beds of such materials, ranging in size from the finest dust to blocks weighing tons, may be built up. Both lavas and fragmental materials are more fully described in Chapter VI.

Both lavas and fragmental materials may build up volcanic cones.

Deformation of Igneous Rocks

Volcanic rocks may, after their eruption, become faulted, jointed, tilted, warped, folded, or foliated in much the same way as sedimentary rocks.

Plutonic rocks may, after their emplacement, become jointed, faulted, warped, or foliated, but large bodies (e.g., batholiths), because of their massiveness and rigidity, are comparatively little affected by folding.

CHAPTER VI

VOLCANOES

Nature and Significance of Volcanoes

A volcano is a vent in the earth's crust out of which hot rocks (either molten or solid) and hot gases erupt. In the popular mind, volcanoes take rank among the most important and real of all geological phenomena. This importance is due to both the terrifying grandeur and mighty power of violent eruptions and their destruction of life and property. Great active volcanoes, like earthquakes, are, however, only relatively minor, outward, sensible manifestations of the tremendous earth-changing forces which operate below the surface. Volcanoes are, from the geological standpoint, much less important than the mighty interior forces which cause the rocks of the earth's crust to be folded and faulted, and large portions of continents to be upraised or depressed. Quantitatively considered, the geological work accomplished by volcanoes is notably less than the simple work of erosion accomplished by running water.

Volcanic eruptions are a surface expression and a limited phase of the major earth process, igneous activity or *vulcanism,* which includes all the activities in the earth's crust which are associated with molten rock and its movement. In our study of igneous rocks we learned that volcanic action is but one of the two important kinds of igneous activity —*plutonic* and *volcanic*—that is, deep-seated shifting and intrusion of molten materials (magmas) into the earth's crust, but not to its surface; and the eruption (or extrusion) of hot rock materials upon the earth's surface. Even as an igneous agency, volcanic action is quantitatively much less important than plutonic (deep-seated) action, although more spectacular.

In making comparisons like those just stated, we must bear in mind the fact that we are dealing with stupendous forces and tremendous masses of the earth's crust, so that volcanic action is, after all, not only a very conspicuous but also a really significant means of changing the face of the earth. The geological importance of vulcanism becomes impressive, indeed, when it is realized that, conservatively estimated, fully

500,000 cubic miles of volcanic rocks have been forced out upon the surface of the earth during the present era of geological time, and that volcanic action was important during each of the five known great eras. In some cases large mountain ranges, like the Cascade Mountains of Oregon and Washington, contain great quantities of volcanic materials.

Volcanic Cones

The accumulation of erupted rock material around the vent causes the building-up of a *cone,* with a pit-like opening at the top, called a *crater,* from which the hot rock materials and gases are ejected. In due course of time the cone may grow to the proportion of a mountain and

Fig. 115. Molten lava seething, boiling, and swirling around an island of solid lava. Crater of Kilauea, Hawaii. (Photo by L. de Vis Norton, courtesy of the National Park Service.)

thus become quite conspicuous. Critically analyzed, the vent in the earth's crust is the *volcano* proper, the process of ejecting the hot materials through the vent is the volcanic *eruption,* and the *cone* is the result, or effect, of the process. Even the greatest of volcanic cones were preceded by simple vents, or fissures, in the earth's crust.

Cones of volcanic origin range in height from a few feet to several miles. Illustrative examples of well-known cones are the following: Mono Craters, California, where there are cones only 10 to 200 feet high; Cinder Cone in Lassen Volcanic Park, California, 640 feet high;

Mt. Vesuvius, Italy, 3880 feet high; Stromboli, about 5000 feet high, as measured from the floor of the Mediterranean Sea on which it stands; Lassen Peak, California (Fig. 140), and Mt. Etna, Sicily, each over 10,000 feet high; Mt. Shasta, California (Fig. 132), and Mt. Rainier, Washington (Fig. 348), each rising to over 14,000 feet above sea level, and 8000 to 10,000 feet above the surrounding country; and Cotopaxi (altitude, 19,600 feet), Chimborazo (altitude, 20,500 feet), and Aconcagua (altitude, 23,000 feet), all of which rise 10,000 to 12,000 feet above the general level of the great elevated platform of the Andes

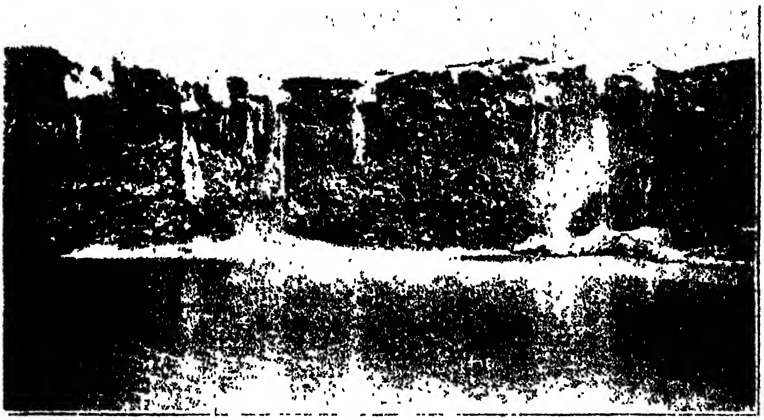


Fig. 116. Molten lava pouring over a cliff into water. Hawaii. (After Diller, U. S. Geological Survey.)

Mountains. Very remarkable cases are Mauna Loa and Mauna Kea on the island of Hawaii, each rising nearly 14,000 feet above sea level, and fully 30,000 feet above the floor of the sea from which they have been built up.

At their bases, volcanic cones are commonly from less than a mile to many miles in diameter. Examples of a few larger ones are: Mt. Rainier, with a basal diameter of over 10 miles; Mt. Shasta, 17 miles; Mt. Etna, 30 miles; and Mauna Loa, with a major diameter of 74 miles, and a minor diameter of 53 miles, measured at sea level. Mauna Loa is probably the biggest volcanic cone on the earth.

Craters of active, and very recently active, volcanoes range in diameter from a few feet to several miles, and in depth from a few feet to several thousand feet. On relatively older, inactive cones, the craters

have of course been partly or completely obliterated by erosion. Very large craters, often called *calderas,* have usually resulted either from violent explosions, which have caused the tops of great cones to be blown away, or by subsidence of the mountain tops. A few examples of craters are as follows: Cinder Cone, California, a few hundred feet wide and 240 feet deep; Lassen Peak, one-fifth of a mile in diameter and a few hundred feet deep; Cotopaxi, one-half of a mile in diameter and 1500 feet deep; Mauna Loa, about two and one-half miles in diameter and 1000 feet deep; Katmai Volcano, Alaska, three miles in diameter and several thousand feet deep; and Mt. Mazama, Oregon (with its Crater Lake), six miles in diameter and several thousand feet deep.

The steepness of the sides of volcanic cones varies from only 5° to 10°, as in the case of Mauna Loa; to 30° or 35°, as in the case of the upper portion of Mt. Shasta; or even to 40° in some cinder cones.

FIG. 117. Wavy, porous lava still hot and steaming. Kilauea, Hawaii.

VOLCANIC PRODUCTS

Hot Gases. Tremendous volumes of gases and vapors are discharged through volcanic vents. The most abundant by far is water vapor, or steam. The quantitative importance of water vapor as a product of great volcanoes may be realized somewhat by consideration of an estimate that about 462,000,000 gallons of water in the form of steam were discharged in 100 days from a secondary cone on the side of Mt.

Etna. Great clouds of steam, usually mingled with more or less volcanic dust, often rise to heights of several miles above large volcanoes during their periods of explosive activity (Fig. 129). Much water

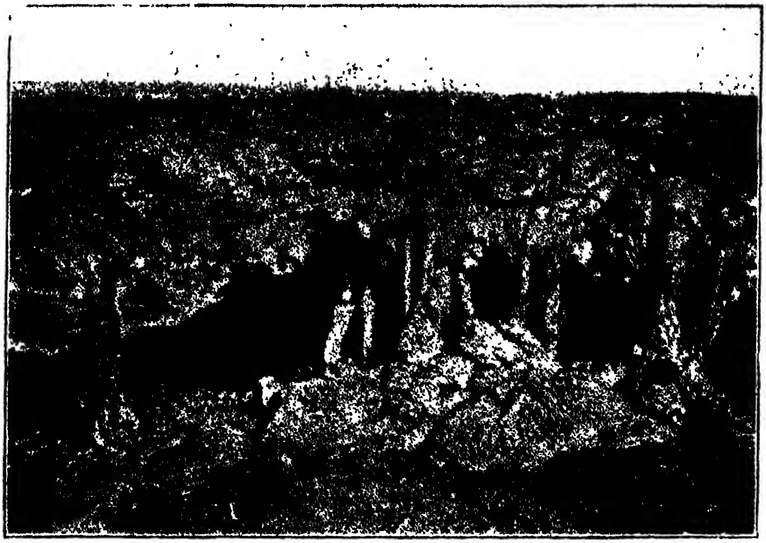


FIG. 118. A lava flow over the edge of an old lava tunnel. Kilauea, Hawaii.

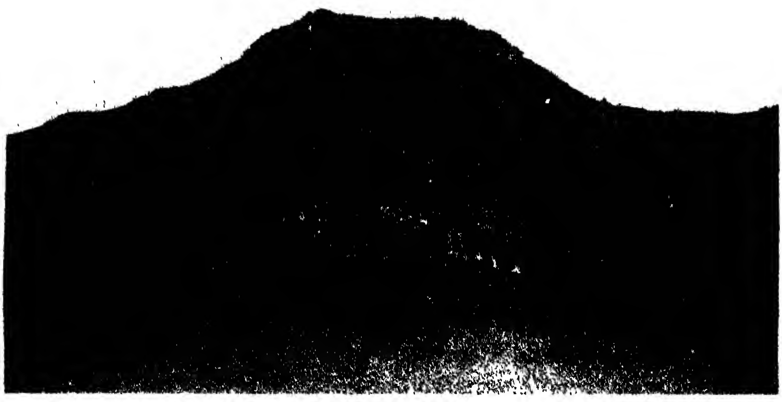


FIG. 119. Lava sheets, representing successive lava flows, exposed by erosion. Near Pahroc Springs, Nevada. (Photo by C. D. Walcott for U. S. Geological Survey.)

vapor also escapes from streams of molten lava, and the discharge often continues for weeks after solidification.

Among the many other gases which are given off by volcanoes and lava flows are the following: sulphide of hydrogen, oxide of sulphur, hydrochloric acid, hydroflouric acid, boric acid, nitrogen, hydrogen, oxygen, and carbonic acid gas. All of these may not be given off during a single eruption or from a single volcano. Some of them may not exist as such in the magmas, because certain chemical combinations may take place immediately after vapors and gases escape into the air before they can be collected and studied.

Lavas. *Lava streams.* The molten materials which issue from volcanoes and fissures in the earth, as well as the rocks which result from their cooling, are called *lavas* (Figs. 115, 116, 117, 118). The source is the hot liquid rock of the earth's interior known as *magma* (Fig. 115). The temperature of magma is very high, commonly ranging from about 1500° to 2500° F. In a general way, increase in the percentage of oxide of silicon (same in composition as quartz) in the various minerals of the magma decreases the temperature necessary to keep the material molten. Increases in gases and vapors (particularly water vapor) in magma also decreases the temperature necessary to keep it molten.

During many volcanic eruptions, magma rises in the crater until it pours over the edge and flows down the side of the mountain in one or more streams, much as would streams of molten iron (Fig. 116). Lava is white-hot when at a high temperature and in a highly fluid condition, but it soon changes to a dull-red glow after it leaves the vent. As the magma flows down the mountain and gradually cools, it becomes a thicker liquid (that is, more viscous), some minerals begin to crystallize in it, and finally the whole

FIG. 120. A spatter-cone on the roof of a lava tunnel. Kilauea, Hawaii.

mass becomes solid lava. A thick lava flow requires months or even years to become thoroughly cooled. Lava streams are very commonly from one-fourth to one-half of a mile wide and from 25 to 100 feet or more deep.

FIG. 121. Detail view of a lava tunnel 40 feet high. Gular, Washington. (Courtesy of the U. S. Forest Service.)

Streams of lava do not always pour out of summit craters of volcanoes. They may break out of the sides of the cones, as has invariably happened in the great active volcano of Mauna Loa, Hawaii, during the last 125 years. In such cases the pressure necessary to lift the columns of molten lava to the summits of the mountains is so great that relief of the pressure takes place by development of one or more fissures on the flanks of the cones out of which the molten lavas pour.

During the process of flowage and cooling of a lava stream, a time comes when there is a strong tendency for a hard, relatively cold crust to form over the still molten material underneath. It is often possible to walk in comparative safety over such crusts. Molten material of a lava stream may, under favorable conditions, drain away under its hardened crust, leaving a long, narrow, more or less winding cave, known as a *lava tunnel*. Such tunnels, which usually range in length from a few hundred feet to several miles, and in diameter from 20 to 50 feet, are often remarkably smooth and regular

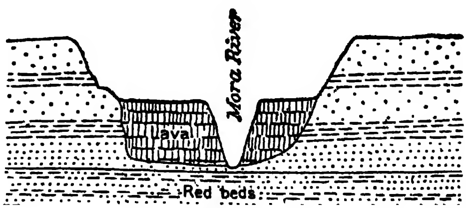


FIG. 122. Structure section showing how a lava flow in a canyon has been trenched by a river. Near Optimo, New Mexico. (After U. S. Geological Survey.)

inside (Fig. 121). Under other conditions, the irregular movement of the lava stream may cause its crust to be broken to pieces, and no tunnel results.

A stream of lava in a very hot, highly fluid condition usually flows down a fairly steep mountainside at the rate of from a few miles per hour to perhaps 10 or 20 miles per hour. As it cools, however, the magma becomes thicker and more viscous, and its rate of motion slowly diminishes until it finally stops. It is not uncommon for streams of lava in Hawaii to continue a slow movement for weeks or even months.

FIG. 123. So-called "pahoehoe" lava at the end of a three-mile flow. Kilauea, Hawaii.

The distance which a lava stream flows is determined by several factors such as temperature, degree of fluidity, kind of molten rock and steepness of slope. Lavas like those of Hawaii and Iceland are of such a nature that they remain fluid at exceptionally high temperatures for so long that they have commonly flowed for 10 to 25 miles, and, in some cases, even 30 to 50 miles. Extreme cases to the contrary are where lavas are so viscous, that is, they have such a low degree of fluidity, that they pile up close around the vents as shown by certain recently extinct volcanoes of France and Germany.

When a stream of molten lava flows through a forest, some of it

may congeal about tree-trunks, the latter being consumed. After the lava flood subsides, hollow tower-like masses of the solidified lava, marking the sites of trees, may remain standing. The so-called "lava trees" of Hawaii are formed in this manner (Fig. 125).

Kinds of lavas. When lavas solidify from a molten condition, they have either a glassy or a stony appearance. Volcanic glass (Fig. 24) (called *obsidian*) is much less common than stony lava. It results from very rapid cooling, especially of the more viscous magmas rich in oxide of silicon. Such a condition is unfavorable for the atoms to build themselves together in the form of crystals, which would give the rock a grained, or stony, appearance. Volcanic glass is, among many other places, finely exhibited in Obsidian Cliff in Yellowstone Park and near Mono Lake; California.

Stony lavas constitute the great bulk of rocks which form from magmas at, and very near, the earth's surface. For their development, the magma must be sufficiently fluid, and time enough must be given during the cooling for crystals (usually small ones) to form. The lava may be wholly crystalline or crystals may be distributed through a glassy groundmass. The mineral composition of some of the most common kinds of lavas—*basalt, andesite, trachyte,* and *rhyolite*—and their relations to other common types of igneous rocks have already been considered.

If, during the consolidation of surface magma, some materials form well-defined crystals scattered through the mass, and then the remaining molten material solidifies with little or no crystallization, a *porphyritic lava* (Fig. 23) results, that is, one with relatively large mineral grains embedded in a much finer grained, or glassy, groundmass.

We have already stated that large quantities of gases and steam often escape from lavas for a considerable time after they are poured out of a vent or a fissure in the earth. Such escape of gases and steam through the upper portion of a lava flow, where the pressure is slight, may fill it with bubbles so that on cooling it becomes *cellular lava* (Fig. 126). If the bubbles are large, giving the rock a spongy appearance, it is called *scoria.* If the bubbles are small, very numerous, and exceedingly thin-walled so that the rock is exceptionally light, it is called *pumice.* In many cases it is light enough to float on water.

Two Hawaiian terms are commonly used to designate the general character of the surfaces of lava flows. One of these is *pahoehoe* which is applied to generally smooth, though often curved and billowy, surfaces of lava (Fig. 123). The other is *aa* which refers to rough, jagged,

badly broken up surfaces, caused either by more or less violent escape of gases or steam or by the breaking up of a hardened crust (Fig. 124) by movement of viscous lava underneath it.

FIG. 124. Part of a rough, badly broken up lava flow. Kilauea, Hawaii.

Fragmental Products. These are the materials (usually heated) which are thrown into the air by the explosive action of a volcano and fall to the ground as solid fragments. They vary in size from the tiniest dust particles to masses of tons weight. The chief sources of such materials are the walls of the throat (or conduit) of the volcano; hardened lava which more or less fills the conduit as a left-over from the preceding eruption; and the upper part of the column of magma which may fill the conduit. Some fragmental products may also be formed by minor explosive action in a stream of molten lava.

Volcanic bombs are pieces of rock, from about an inch to several feet in diameter, which are hurled out of volcanoes. They may be more or less angular blocks torn loose in solid condition, or they may result from violent disruption of molten material whereby masses of the magma, in whirling through the air, take on somewhat rounded forms and solidify as such. Bombs of the latter sort are often cellular (Fig. 127) or even pumiceous because of the escape of gases.

Volcanic cinders are fragmental materials ranging in size from about

an inch down to dust particles. Larger cinders are called *lapilli,* and smaller ones are called *volcanic ashes.* Both types may or may not be porous. The terms "cinders" and "ashes" are good only in the sense that they suggest a resemblance to familiar products of burning, but they are not results of combustion. More or less well-defined beds or layers of the larger fragments (blocks, bombs, and cinders) produced by successive eruptions, and cemented with ash or other substances, form *volcanic breccia* (Fig. 25).

FIG. 125. A so-called "lava tree." Island of Hawaii.

Volcanic dust is the most finely divided material ejected from volcanoes. It may be so finely pulverized as to be an impalpable powder which may be sent miles into the air to remain suspended for weeks or months and be carried by atmospheric currents for hundreds or even thousands of miles. The eruption of dust is an important part of the geological work of volcanoes. Close around the vents of explosive volcanoes, dust not uncommonly accumulates to depths of many feet and, 50 to 100 miles away, to depths of several inches. Successive eruptions often cause volcanic dust and ashes to accumulate in more or less well-defined layers or beds which become compacted into a rock called *tuff.*

TYPES OF VOLCANIC ERUPTIONS

Volcanoes are conveniently classified into three principal types, namely, (a) effusive or *quiet,* (b) *explosive,* and (c) *intermediate.* This is a natural classification based upon the mode of eruption, involving characteristics of the eruptive activity and the kinds of materials ejected. The latter, in turn, determines the style of the cone which is built. Volcanoes are frequently referred to as *active, dormant,* or *extinct.* This nomenclature refers to the *state* of activity rather than to the characteristics of eruption and may be applied to any of the three types of vol-

canoes as classified above. For purpose of classifying volcanoes these terms (active, dormant, extinct) are not satisfactory because, as has frequently happened, a volcano which has been inactive for many years may again break forth. Vesuvius in Italy and Lassen in California are among many examples.

In addition to the three modes of eruption just mentioned, there are also fissure eruptions and domal eruptions.

Effusive or Quiet Type. Volcanoes characterized by effusive eruptions are relatively quiet in action and comparatively free from severe explosions. Streams of lava either well up in their craters and overflow their rims or break out of the flanks of the mountains and flow down their sides. The lavas of such eruptions are usually in a highly fluid condition and flow for miles. Gases and steam of course escape from them in large quantities, but rarely, if ever, with great violence. The two great active volcanoes of Hawaii—Mauna Loa and Kilauea—are excellent examples of the effusive type. They are briefly described later in this chapter.

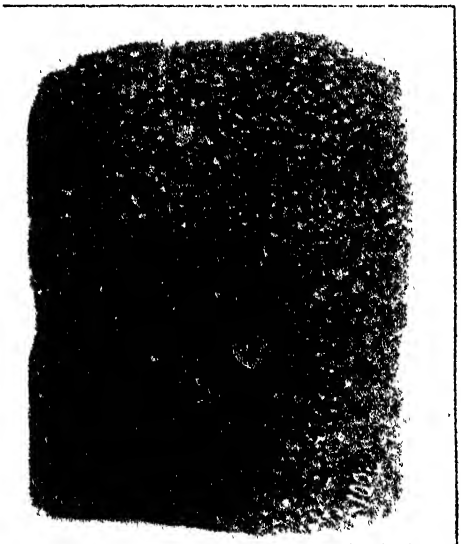


FIG. 126. A specimen of cellular lava.

Volcanic cones built up wholly or largely by many effusive eruptions are generally characterized by having large craters (or calderas), low angles of slope (usually less than 10°), and great basal diameters. The two last-named features are due to the fact that the lava streams tend to flow far out from the vents. Such cones are spoken of as *shield cones* or lava cones.

Explosive Type. Volcanoes characterized by explosive eruptions are violent in action. In extreme cases the top of a cone or even almost the entire cone may be blown to pieces and widely scattered. An example of extreme violence was that of Krakatoa in 1883, described

later. Typical explosive volcanoes seldom yield lava streams, but they commonly send great clouds of volcanic dust and ashes, mingled with steam, high into the air. Large blocks of rock are also often hurled out. Some weigh as much as several tons.

Volcanic cones built up largely by explosive eruptions generally have well-defined craters; their sides are steep (up to 30° or 40°); and the diameters of their bases are relatively small. The two last-mentioned features are due to the fact that most of the materials, in solid form, particularly the coarser fragments, accumulate relatively close around the vents and produce slopes much steeper than lava flows. *Cinder cones*, built up by explosive eruptions of volcanic cinders, belong in this category (Fig. 130).

Intermediate Type. Most of the volcanoes of the world, especially the larger ones, are neither typically effusive nor explosive in action but rather intermediate between the two. They are characterized by more or less alternation of eruptions of lava streams and fragmental materials. Mt. Shasta, California, and Mt. Rainier, Washington, are the two greatest volcanic cones of the intermediate type in the United States. The active Mt. Vesuvius is another good example. Such cones are usually rather steep-sided, that is, their slopes are often 20° to 30° and are referred to as *composite* cones.

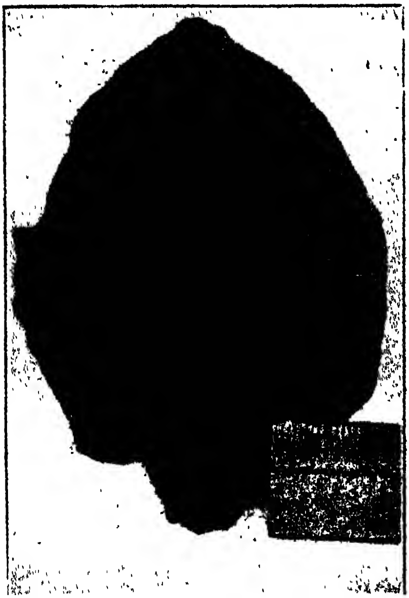


Fig. 127. A volcanic bomb from southern Idaho.

Fissure Eruptions. It has already been suggested that not all volcanic materials are erupted through cones. Eruptions may take place through fissures, both small and great, in no way connected with volcanoes in the ordinary sense of that term. The materials thus erupted are always molten rather than fragmental. Some of the best exhibitions of fissure eruptions during the last century and a half have been those of Iceland.

Thus in 1783 molten lava poured forth from many places out of a fissure 20 miles long. One of the streams of lava was nearly 50 miles long, and another nearly 30 miles long. Each was several miles wide. As late as 1913, molten lava welled out of a number of very small craters arranged along a fissure and spread out over the plains.

Fissure eruptions have in past ages produced vast fields of lava of great depths. Thus, much of the Columbia Plateau, covering about 125,000 square miles in Washington, Oregon, Idaho, and northeastern California, has been built up of successive flows, mainly from fissures, to a depth many hundreds to several thousand of feet. Valleys were filled, hills were buried, and some mountains were surrounded by the molten floods. These eruptions occurred during the present (Cenozoic) era.

FIG. 128. Paricutin, the new volcano 200 miles west of the City of Mexico, shown during an eruption in January, 1944, when the cone was over 1500 feet high. Eruptions began on flat, cultivated land in February, 1943. (Photo by I. C. McBride, courtesy of M. R. Huberty.)

Domal Eruptions. When a magma is too viscous to flow, it may be forced out of a vent to form a steep-sided, more or less dome-shaped or pluglike mass, called a *volcanic dome,* as in the case of Lassen Peak, California. The temporarily existent spine of Mt. Pelée was in this category.

Age and History of Volcanoes and Volcanic Cones

New Volcanoes. A considerable number of relatively small volcanic cones are known to have been built up during the Christian era. Some of these have developed on land, and some in the sea, forming islands. A few examples will be given.

Monte Nuova, a cone 440 feet high near Pozzuoli, Italy, was built up in 1538. A vent was formed by bending up and breaking the

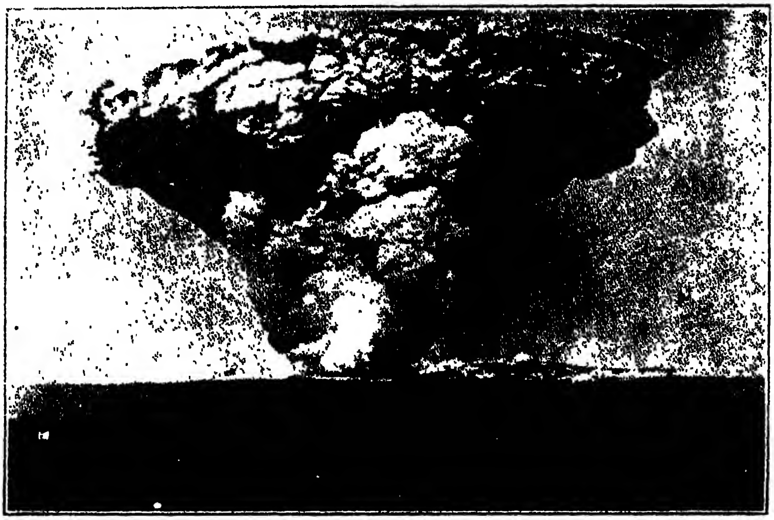


Fig. 129. The grand eruption of Lassen Peak, California, May 22, 1915. The volcanic cloud was fully 8 miles high. Photo taken at Anderson, 50 miles away. (Photo by courtesy of Myers and Loomis.)

ground. Glowing lava was visible, and eruptions of fragmental materials continued for about a week, building up the cone. There have been no eruptions since. The cone stands among others which are not much older.

A remarkable case is that of Jorullo, Mexico, where a volcano burst forth in cultivated fields one day in 1759. Eruptions continued for five months. Large quantities of both molten and fragmental materials were ejected, building up a cone to a height of about 1000 feet. A little later (in 1770), activity started at Izalco in San Salvador. Eruptions, often violent, have been almost continuous since that time, and a cone over a mile high has been formed.

Cinder Cone (640 feet high) and its associated lava field of several square miles came into existence in northeastern California as a result of eruptions which began not longer ago than the early part of the sixth century. The second and last lava flow, which poured out during the middle of the nineteenth century, is probably the youngest lava in the United States.

In 1831 vigorous volcanic activity on the floor of the Mediterranean Sea caused an island of fragmental material 200 feet high to be formed. In a relatively short time it was cut away by wave erosion

In the Santorin Islands of the Greek Archipelago, several small islands have been formed by volcanic activity during the last 2000 years.

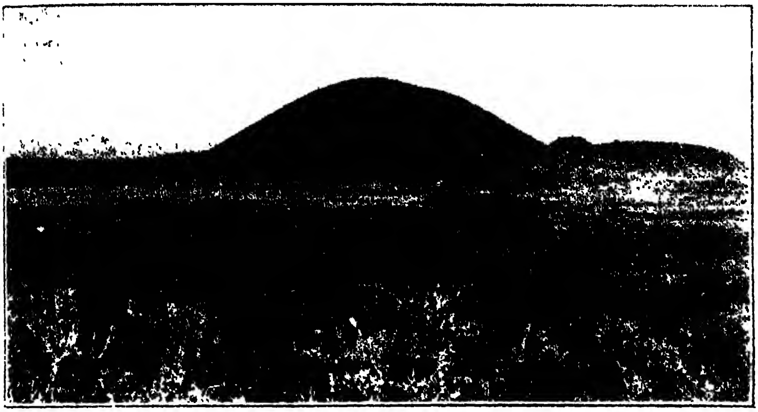


FIG. 130. Group of recent cinder cones. San Francisco Plateau, Arizona. (Photo by Gilbert, U. S. Geological Survey.)

A number of spectacular eruptions in the Aleutian Islands of Alaska, particularly in 1796, 1883, and 1906, have resulted in the formation of islands in the sea.

Various cinder cones in Arizona (Fig. 130) and eastern California are so fresh and unaffected by erosion that they certainly cannot be more than a few hundred, or at most a few thousand, years old.

On February 20, 1943, volcanic activity began at Paricutin about 200 miles west of the City of Mexico. Paricutin volcano is of special interest because its activity has been observed by scientists almost constantly since the day of its birth. Eruption began on level, cultivated land. At first hot fumes appeared, followed quickly by fragmental materials thrown into the air. Soon there were lava flows, and more

Fig. 131. An aerial view of recent cinder cones and a lava flow. The lava flow, five miles long, emerges from the base of a cone and spreads out. About forty miles southeast of Grand Canyon, Arizona. (Fairchild Aerial Surveys.)

fragmental materials were hurled out. Within ten weeks two volcanic cones—one of them 1100 feet high—were built up. Many earthquakes occurred during a period of 20 days preceding the first eruption, and many more accompanied eruptions until May by which time activity had subsided considerably. Then there was a strong earthquake followed by renewed activity accompanied by lava flows. All farms and villages within several miles were abandoned. The village of Paricutin was partly covered by lava. One of the cones reached a height of 1500 feet by September, 1943. According to reports in 1947, Paricutin was still active.

A spectacular example of submarine volcanic activity, resulting in the building of a small island in the Pacific Ocean, occurred early in 1946 about 200 miles south of Yokosuka, Japan.

Duration of Volcanic Activity. The length of time during which individual volcanoes remain more or less active is exceedingly variable, ranging from a few days (or less) to hundreds of thousands, or even millions, of years Most of the great volcanoes of the present time began their activity in the latter part of the present (Cenozoic) era. They are, therefore, several million years old. Some of them, like Kilauea, are now constantly active; some, like Mauna Loa and Mt. Etna, are very active at intervals of a few years; others, like Mt. Shasta and Mt. Rainier, are either dormant or practically extinct; while still others, like Mt. Crandall in Yellowstone Park, ceased action so many hundreds of thousands of years ago that the great cones, many miles in diameter, have been very largely removed by erosion. We may gain some conception of the age of big individual cones when we realize that one like Mt. Etna has, in spite of many great eruptions, remained practically unchanged in its general outline for more than 2000 years.

The evidence is plain, from the study of historical geology, that volcanic activity took place during the earliest known (Archeozoic) era of earth history, and that such activity has occurred on small and grand scales in many parts of the world, and during various periods, since the earliest known time.

Destruction of Volcanic Cones. In some cases volcanic cones are partly or almost wholly destroyed through their own explosive activity. Thus an explosion of terrific violence in Katmai Volcano, Alaska, in 1912 blew away several cubic miles of the top of the mountain (Fig. 137), and the explosion of Krakatoa in the East Indies in 1883 almost completely obliterated what was a fair-sized cone. A cone may be

partially destroyed by engulfment or subsidence of its upper portion due to weakening of the support underneath. The great crater (or caldera) of Mt. Mazama, containing Crater Lake, in southern Oregon was thus formed (Fig. 359).

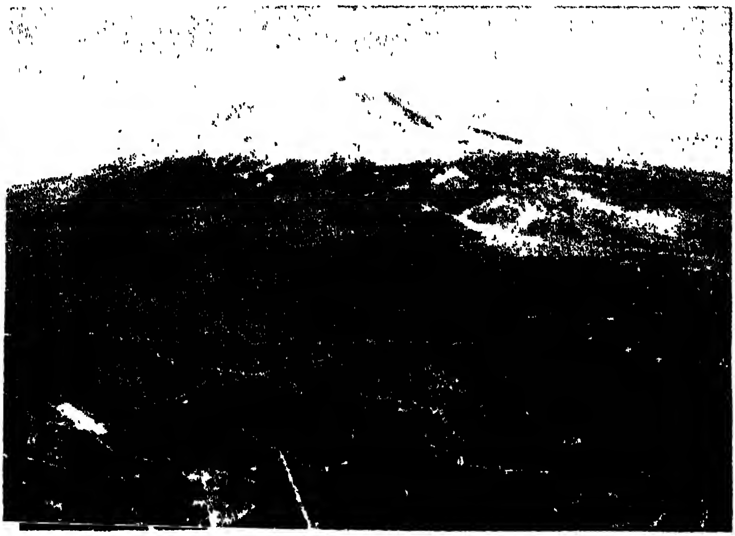


FIG. 132. A great, considerably eroded, volcanic cone rising about 10,000 feet above the surrounding country. Mt. Shasta, California. (Fairchild Aerial Surveys, Inc.)

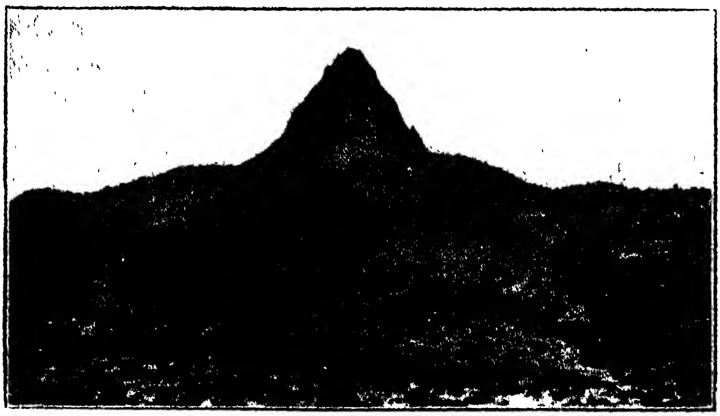


FIG. 133. A volcanic neck. Mt. Taylor region, New Mexico. (After Dutton, U. S. Geological Survey.)

The destructive work of weathering and erosion is, however, the greatest cause of obliteration of volcanic cones Every volcanic cone, even when in course of construction, is subjected to the attacks of weathering and erosion. The upper part of the cone of Mt. Vesuvius was, for example, distinctly trenched by erosion soon after the great eruption of fragmental materials over its sides in 1906. When, barring very violent eruptions, the amount of material ejected by a volcano is greater than can be removed by erosion, the cone continues to build up.

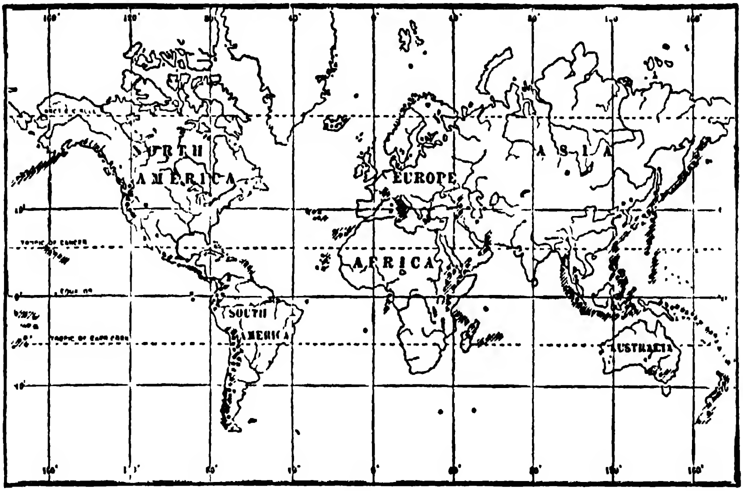


FIG. 134. Map showing the distribution of active and recently extinct volcanoes. (From Tarr's *New Physical Geography*, by permission of the Macmillan Company.)

The activity diminishes and finally ceases, after which the cone becomes more and more deeply dissected (Fig. 132), its crater becomes obscured, and its height gradually becomes lower. During a late stage of its erosion, nothing but the core or plug of the volcano may rise above the general level of the country (Fig. 133), and finally it may completely vanish as a topographic feature. If, however, the more or less dissected cone should become buried in the earth under accumulations of sedimentary rocks, it might, at a much later time, again become exposed at the surface. An interesting case in point is a very ancient volcanic landscape which is now again coming to light as erosion proceeds in a part of Great Britain.

Rebuilt Volcanic Cones. When a cone is subjected to a great catastrophe such as a violent explosion or a profound subsidence of its upper part, a crater pit (or caldera) of large size usually results. In the case of Katmai Volcano, already mentioned, the explosions of 1912 left a hole several miles in diameter and several thousand feet deep. The cone has not even been partially rebuilt since the catastrophe, although it may be in the course of time.

The vast caldera, over five miles wide and several thousand feet deep, which resulted from destruction of the upper part of the cone of Mt. Mazama in Oregon some thousands of years ago, has been only very slightly rebuilt by volcanic activity. Wizard Island (Fig. 359) is a product of such subsequent activity.

A large portion of Mt Vesuvius was blown away during the great eruption of A.D. 79. Eruptions since then have built the present cone upon the stump of the old mountain.

Submarine Volcanoes

It has already been suggested that volcanic activity may take place on the floor of the sea. A remarkable example is Mauna Loa, Hawaii, which began action at the bottom of the mid-Pacific Ocean where the water was fully three miles deep. It has been built up into a gigantic, gently sloping cone nearly 14,000 feet above sea level. All the eight Hawaiian Islands mark the exposed portions of a great submarine volcanic ridge or range several hundred miles long.

A remarkable case of a great mountain range being built up out of the sea is the chain of Aleutian Islands, Alaska, more than a thousand miles long. It contains various active volcanoes—three new ones (the Bogoslov volcanoes) having been built up as islands in the years 1796, 1883, and 1906.

The eastern portion the West Indies is of very recent submarine volcanic origin, with certain volcanoes, like Mont Pelée and La Soufrière, still active. The East Indies are also to a considerable extent of volcanic origin, with numerous active cones.

Among many other examples of volcanoes of submarine origin, mention may be made of the Azores, Cape Verde Islands, Canary Islands, and various islands of the south Pacific Ocean. Mention has already been made of the cone (Graham's Island) which was built up by eruptions in the midst of the Mediterranean Sea in 1831.

Distribution of Active and Recently Active Volcanoes

Hundreds of volcanoes are definitely known to be active, and several thousand others are either dormant, or have become extinct in very recent geologic time, that is during the latter portion of the present era. Most of these volcanoes show a strong tendency toward arrangement into two grand zones or belts (Fig. 134). One of these belts nearly encircles the Pacific Ocean, extending through western South America, Central America, western North America, the Aleutian Islands, Kamchatka, Japan, the Philippine Islands, the East Indies, the New Hebrides, and New Zealand. There are, of course, various local portions of this belt without volcanoes. The other great belt is less well defined and more interrupted. Beginning, let us say, in Central America, it extends through the eastern part of the West Indies, the Azores, the Canary Islands, the Mediterranean region, Asia Minor, southern Arabia and eastern Africa, eastern India, the East Indies, and the Hawaiian Islands. A considerable number of volcanoes lie outside of the two grand belts.

Various ideas have been expressed in the attempt to explain the distribution of most of the active and recently active volcanoes in the two great belts. Without entering into this discussion, suffice it to say that these volcanoes occur in zones where earth-crust disturbances have been recently, and are now, unusually pronounced. They are, in other words, in zones of exceptionally active mountain-building movements. These zones are, in a general way, also the belts of greatest earthquake activity, and, as already stated, both earthquakes and volcanoes are but surface and near-surface manifestations of deeper-seated and more profound earth-crust activity.

Examples of Volcanic Eruptions

A few examples of volcanic eruptions will now be described briefly in order to give the reader a still more definite conception of the degrees of violence of eruptions, from relatively quiet to highly explosive; of the types of eruptions; the kinds and quantities of materials erupted; the tremendous power involved; and the manner in which volcanic cones may be altered by eruptions.

Mauna Loa and Kilauea. Two of the most interesting, readily accessible, great volcanoes are Mauna Loa and Kilauea on the island of Hawaii in the midst of the northern Pacific Ocean. They are fine illustrations of

the effusive or relatively quiet type of volcano. Mauna Loa is an exceedingly large volcanic pile with very gently sloping sides rising to nearly 14,000 feet above the sea, and Kilauea lies on its flank at an altitude of about 4000 feet. Each has a vast, oval-shaped crater about three miles long, bounded by nearly vertical walls of lava many hundreds of feet high. Each crater pit has a nearly level floor consisting of hard, fresh, black lava which is really only a crust covering a mighty column of molten lava, several miles in diameter, extending far down into the mountain. An eruption of Mauna Loa is often preceded by earthquakes. It usually begins with lava breaking

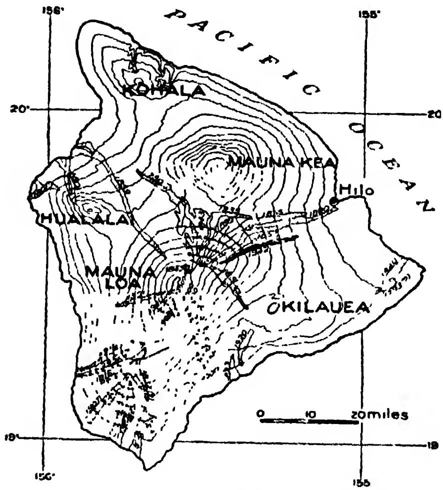


FIG. 135. Dated lava streams on Mauna Loa. (After G. A. Macdonald.)

through the crust in the great crater pit in the form of large and small fountains. During the last 100 years or more, lava has not overflowed the crater rim. Instead, one or more streams of molten lava break out somewhere on the side of the mountain several thousand feet from the summit. Fountains of the magma, at the vents, often rise 100 to 300 feet. Many resulting lava streams, one-fourth of a mile to a mile wide, have flowed down the sides of the mountain 10 to 45 miles, sometimes even into the sea (Fig. 135). The great lava stream of 1919 entered the sea and poured into it for weeks, after flowing about 15 miles from the source on the flank of the mountain (Fig. 136). Important eruptions occur at intervals of about 2 to 7 years. Between such eruptions, Mauna Loa shows little evidence of activity. Recent eruptions occurred in December, 1933, November, 1935, and April, 1942.

Kilauea acts in general much like Mauna Loa. Lava streams have poured out of its flanks also at various times, in each case preceded by a

rise of the lava column in the vast crater pit. Within the mighty crater bowl of Kilauea there is, however, an inner pit or crater, known as Halemaumau, about three-fifths of a mile in diameter marking a place where the crust of the great column of molten lava in the throat of Kilauea is broken through (Fig. 115). Until 1924, within this nearly circular inner pit with vertical walls, the incessantly molten lava rose and sank hundreds of feet within periods of a few years. Sometimes the magma overflowed the pit and streamed out upon the wide floor of Kilauea (Fig. 117). It was,

FIG. 136. The great lava stream of 1919 entering the sea after flowing 15 miles down the side of Mauna Loa, Hawaii. The high temperature of the molten lava caused a tremendous commotion in the sea, accompanied by great steam clouds. (Photograph furnished by Bessie Wood, Northampton, Massachusetts.)

indeed, an awe-inspiring sight to look into the inner pit of Kilauea, especially at night. "The boiling lava is apparently white-hot at a depth of but a few inches below the surface, and, in the overturnings of the mass, these hotter portions are brought to the surface and appear as white streaks marking the redder surface portions. From time to time the surface freezes over, and the cracks open and erupt at favored points along the fissures, sending up jets and fountains of lava, the material of which falls in pasty fragments" (W. H. Hobbs). For the past several years the lava has been frozen over and stands as a cold, black floor about 800 feet beneath the rim of the pit.

Krakatoa. Three examples—Krakatoa, Katmai, and Pelée—will be described to illustrate highly explosive volcanic activity. Among these, Krakatoa, a volcanic island between Sumatra and Java, had been dormant for over 200 years. Then, in August, 1883, a series of terrific explosions lasting two days caused more than a cubic mile of rock material to be thrown into the air in the form of volcanic pumice, ashes, and dust. The site of the island was mainly covered with water 600 to 900 feet deep immediately after the catastrophe. Some of the most violent explosions were heard hundreds of miles away. The atmosphere of the whole world was disturbed, as recorded by rise and fall of barometers. Part of the vast cloud of dust rose fully seventeen miles into the air. Dust fell in perceptible amounts over an area of several hundred thousand square miles. Small quantities of the finest material filled the whole earth's atmosphere, remaining suspended for months and causing the famous red sunsets of the fall and winter of 1883–1884. Sea waves 75 to 100 feet high, caused by the disturbance, rushed upon the neighboring coasts of Java and Sumatra and killed 40,000 people.

FIG. 137. The remains of Katmai Volcano, Alaska, after the great eruption of 1912. The whole upper part of the mountains was blown away by the explosion, leaving a hole more than 2 miles in diameter and several thousand feet deep. (Copyright the National Geographic Society. Reproduced by special permission.)

Katmai Volcano. Within two days in June, 1912, Katmai Volcano in southern Alaska was subjected to several terrific explosions which were probably of even greater violence than those of Krakatoa. The cone, which rose over a mile above the surrounding country, had its whole upper portion, involving about 5 cubic miles of rock, blown away, leaving a vast crater (or caldera) two and one-half miles in diameter and several thousand feet deep

(Fig. 137). This crater is now one of the world's largest. The first and greatest of the three explosions was heard in Juneau, Alaska, 750 miles away. The product of the explosions was mainly dust which fell to a depth of one foot in a village 100 miles away, and in perceptible amounts 900 miles away in southeastern Alaska. Dust and larger fragments fell to depths of 2 to 10 feet on the flanks of the beheaded mountain. Glaciers on the mountain were truncated, leaving walls of ice over two miles long capping part of the crater rim. Severe earthquakes accompanied the explosions.

Mont Pelée. Mont Pelee, situated on the island of Martinique in the eastern part of the West Indies, was violently active in May, 1902. The last eruption prior to that was in 1851. For a few weeks before May 8, 1902, there was considerable activity accompanied by earthquakes, but on that date a great explosion caused a tremendous cloud of hot gases filled with incandescent particles of dust to rise out of the crater. Because of its weight, this vast cloud rushed with hurricane velocity down the side of the mountain. The city of St. Pierre lay in the path of the descending, fiery cloud, and its whole population of nearly 30,000 people (excepting one or two persons) was almost instantly annihilated. Fire finished the work of destruction. A number of violent eruptions took place within the next few months. In October, a remarkable feature began to develop in the crater at the summit of the mountain. This was a great column of steaming-hot, more or less pasty lava which slowly rose and solidified, reaching a height of about 1000 feet in seven months (Fig. 138). It gradually crumbled to pieces.

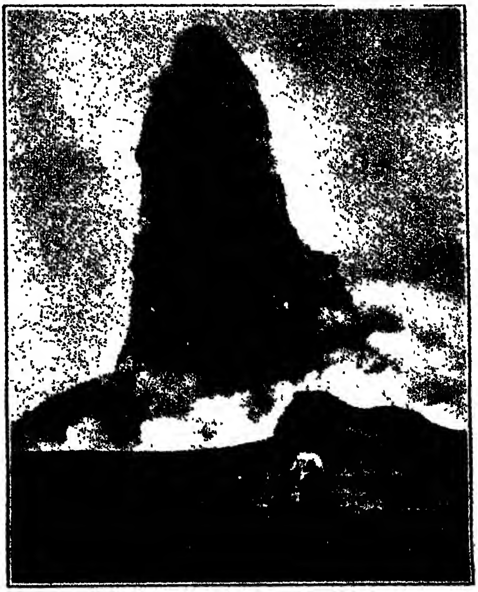


FIG. 138. The great spine at the summit of Mt. Pelée in the West Indies in 1902. The steaming hot spine was about 1000 feet high. (Photo by E. O. Hovey, courtesy of the American Museum of Natural History.)

Mt. Vesuvius. Excellent examples of volcanoes of the intermediate type are Vesuvius and Etna. For centuries prior to the Christian era, Mt. Vesuvius seems to have been inactive. From A.D. 63 to 79, nu-

merous earthquakes shook the mountain and vicinity. Then, in the year 79, there occurred the most violent eruption of the mountain in historic times. No molten lava appeared, but the explosion blew away much of the upper part of the cone, greatly altering its outline, and leaving a conspicuous crescent-shaped ridge around part of the stump of the mountain. "Ashes fell upon the surrounding country, a huge column of steam and ash darkened the sky, and great torrents of water fell upon the flanks of the mountain. Pompeii was buried beneath a cover of ash and dust which penetrated every crevice, and so sealed the objects in a compact cover. In the excavations which have been made during the last century, objects of even a perishable nature have been recovered. . . . It is a wonderful experience to walk through the deserted streets of this ancient city of 20,000 inhabitants, to realize under what terrible conditions the people were driven out or overwhelmed in their efforts to escape" (Tarr and Martin). The city of Herculaneum was, at the same time, overwhelmed by a great flow of hot mud, formed by clouds of condensing steam mixed with ashes. Between A.D. 79 and 1139, a number of eruptions occurred. Then for nearly 500 years there was scarcely any activity. One of the greatest eruptions of Vesuvius occurred in 1631 when large quantities of ashes and dust were ejected, and several streams of molten lava poured out of the crater and down the sides of the mountain, overwhelming some villages. Since that time activity varying from mild to vigorous has been almost continuous, and the present cone (altitude nearly 4000 feet) has been built upon the stump of the mountain left by the explosion of the year 79. A grand eruption took place in 1872 when vast clouds of ashes and dust were thrown high into the air, and streams of lava flowed out of fissures in the sides of the cone. A later eruption, nearly as great as that of 1872, took place in 1906 when ashes and dust were thrown miles into the air, and several streams of lava flowed out of breaches in the mountainside.

Mt. Etna. The cone of Mt. Etna in Sicily rises to a height of nearly 11,000 feet. Its base is over twenty-five miles in diameter. Many vigorous eruptions have taken place since the first known one in 476 B.C. During the last 100 years, eruptions have occurred at average intervals of about 5 years. Destructive earthquakes usually precede and accompany the violent eruptions. A typical eruption is characterized by a series of explosions which send vast clouds of ashes and steam into the air from the great summit crater, while, from several (or many) openings on the flanks of the mountain, there issue streams of molten lava, some of which flow down to the base of the great cone, and even into the sea. The openings from which the lava streams emerge are secondary craters, often in small cones, hundreds of which occur on the sides of Etna. Grand eruptions took place in 1910–1911 when many vents on the sides of the mountain poured forth lava,

and the summit crater ejected dust and ashes. The latest great eruption took place in 1923.

FIG. 139. A great volcanic cloud which has been called "Vulcan Face." Lassen Peak, California. (Photo by B. F. Loomis.)

Lassen Peak, California. In conclusion, brief mention may be made of Lassen Peak in northern California which is of special interest not because of the magnitude of eruptions, but because it is the most recently active volcano in the United States. The steep-sided cone of Lassen rises about a mile above the surrounding country (Fig. 140). Prior to May 30, 1914, the mountain had been inactive for hundreds or even thousands of years, as judged by the state of weathering of its crater. On the date mentioned, the old volcano suddenly burst into explosive activity, and hundreds of eruptions occurred within the next few years. Little or no lava appeared, but great clouds of steam and dust were shot into the air, often to heights of several miles (Fig. 129 and 139) and scattered ten to thirty miles around the mountain. During a grand eruption of 1915, a tremendous volume of condensing steam mingled with volcanic dust started down the eastern face of the cone, causing the snow to melt. The resulting flood of hot mud and loose rock fragments, together with the very hot volcanic cloud, rushed with terrific speed to the base of the cone and into a beautiful mountain valley, leaving an appalling scene of desolation for ten miles. Forests were swept away for miles, and fires were set (Fig. 141). Real eruptions ceased during 1917, but some steam still escapes within the crater. Renewed activity may occur.

CAUSE OF VOLCANIC ACTIVITY[1]

The problem of the cause (or causes) of volcanic activity is one of the most uncertain in geological science. Our present purposes are to call attention very briefly to some of the more important facts involved and to offer a few explanatory suggestions.

[1] This statement of the cause of volcanic activity is taken essentially from the author's volume 3 of *Popular Science Library* published by P. F. Collier and Son Company.

A long-held idea that a relatively thin crust covers a molten earth-interior, and that downward pressure of this crust, due to earth contraction, causes molten rocks to be forced out, has been too thoroughly disproved to be now seriously entertained. The fact that near-by volcanoes, like Mauna

Fig. 140. A midsummer view of Lassen Peak, California, from the east, before the eruptions began in 1914. (Photo by B. F. Loomis.)

Fig. 141. Lassen Peak as it appeared from the east after the devastating eruption of 1915. Compare with Fig. 140. (Photo by A. L. Day, Carnegie Institution of Washington.)

Loa and Kilauea, commonly erupt entirely independently shows that there can be no universal liquid beneath a relatively thin crust. Other arguments against liquidity of the earth's interior are that the earth acts like a body nearly as rigid as steel against the powerful tide-producing forces, and that earthquake waves, which pass through the earth to a depth of at least 2000 miles, are of the kind which require a solid medium for transmission.

We may, then, consider more plausible views in regard to the cause of vulcanism. First of all, we may be sure that the earth is highly heated inside. Measurements made in deep borings show that the temperature increases downward at the rate of about 1° F. for each fifty to seventy-five feet to depths at least as great as several miles. The temperature must, therefore, be several thousand degrees at depths of twenty-five to forty miles. This is sufficiently high to cause all ordinary rocks to melt at the earth's surface. At great depths, however, the downward pressure on the rocks is so tremendous that their melting points are notably raised, so that there is every reason to believe that the rocks twenty-five to forty miles down are in general not molten.

If we adhere to the older (nebular) hypothesis of earth origin, the interior heat of our planet is left over from its once molten condition. On the basis of another (planetesimal) hypothesis, the earth's heat is maintained by the steady, powerful action of gravity, which causes the earth to contract. The earth is, in any case, hot inside as proved by deep-well records and by igneous phenomena in general, and it is a shrinking body as proved by the many large-scale zones of wrinkling and folding of the rocks. If, then, highly heated, solid rocks at reasonable distances down in any part of the earth are subjected to relief of pressure by an earth movement, such as upward crumpling or bending of the crust, or by readjustment of large fault blocks, such heated, solid rocks may become locally molten. The same crustal disturbance which brings about such relief of pressure and melting may very reasonably be regarded as the power which forces some of the newly formed molten material higher up into the crust and even out upon its surface. This view harmonizes with the well-known fact, already mentioned, that the main belts of active volcanoes are also the main belts of active earth movements, such as earthquakes.

Another source of power behind volcanic action is steam and gas pressure. We have already referred to the fact that tremendous amounts of water in the form of steam escape from volcanoes and even from streams of molten lava. The violent volcanic explosions are all, or nearly all, direct results of giving way of volcanic cones to steam (and gas) pressure which increases during greater or less periods of time, and with little or no possibility of escape without rupturing the mountain. Steam alone or combined with some other gases may also aid in forcing out the molten rock.

What is the source of the steam and other gases or vapors? According to one view, they were originally within the earth. According to another view, the water, at least, has been absorbed by the molten rocks from surface waters which have worked their way downward into the earth's crust. At least three arguments are opposed to the second hypothesis: first, that a considerable number of volcanoes are many miles from the sea or other large bodies of water; second, that downward percolation of rain water

would fall far short of supplying the tremendous quantities of water ejected by volcanoes; and third, that any water taken up by molten rock must be absorbed within a very few miles of the surface because farther down (in the zone of flow) there are no openings large enough to permit any very notable downward passage of water. As a matter of fact, the uppermost portion of the earth's crust is just where magmas give up their water, often with great violence.

CHAPTER VII

ROCK WEATHERING

Introduction

The Place of Weathering. All the materials of the outer, or crustal, portion of the earth are subject to ceaseless change. Nothing endures. Under the action of the weather and other more or less closely related agencies, even the hardest and most resistant rocks crumble or decay in the course of time (Figs. 142, 149, 151, and 156). It has already been pointed out that *gradation* is the most significant of the three major earth processes which fall in the realm of physical geology (Chapter I). It includes *all* activities which work upon the earth's *surface* to reduce its irregularities to a common level. The many agents working toward such a general end include the atmosphere, wind, running water, glaciers, and others. Some of the processes which are carried on by these agents are destructive, that is, they tend to break down the rocks and to reduce the earth's surface to lower levels. Other processes are constructive and tend to build up the lower places through the deposition of materials which are transported from higher levels. Thus, there are two main subdivisions of gradation, called *degradation* (the destructive processes) and *aggradation* (the constructive processes).

Weathering is one of the partners in degradation. The term refers to the natural processes of disintegration and decomposition of rock. It includes, therefore, those activities whereby the rocks at or near the earth's surface break up, decay, or crumble. Since the weathered materials (rock residue) generally remain where they were formed, until subsequently removed by agents of erosion, it is essentially a static process. Soil is the end-product of rock weathering. In the general work of degradation a companion process to weathering is *erosion*. It is the natural removal and transportation of rock material. It requires a moving agent, such as running water or wind, which picks up rock material and moves it elsewhere. It is a dynamic process. Weathering prepares the rock for erosion, although in some instances solid rock is eroded direct (by scouring) without the intermediate, loosening activities of weathering.

Effects of weathering are, as a rule, scarcely noticeable during the ordinary span of a human life, but, during the eons of geological time, weathering processes have been relentlessly at work upon the surface and near-surface portions of the earth, causing such tremendous quantities of rock material to be broken up and decomposed that the lands have been profoundly affected. In fact, most of the materials by far which make up the vast bodies of sedimentary rocks are products of rock weathering which have been transported from their places of origin.

Rate of Weathering. The rate of weathering depends upon the nature of the rocks and the kinds and conditions of the weathering agents which operate upon them. It is a matter of common knowledge that many stone buildings and monuments show marked effects of weathering. An excellent case in point is Westminster Abbey in London which was built of weak, rather porous stone in the thirteenth century. Many of its outer stones are badly weathered, some of its ornamental, carved parts having been reduced to shapeless forms. Many of the exterior carvings of soft limestone of the Louvre in Paris are also badly weathered. Inscriptions on many tombstones and monuments only one or two centuries old are often nearly or quite illegible, due to weathering. Weathering is, however, much less rapid in the case of hard, resistant rocks. Thus, even the polished surface of a very resistant rock, like granite or quartzite, may be preserved for many years although exposed to very vigorous and changeable weather conditions. There are ways of estimating that many of thousands of years are required for enough weathering to develop a soil a few feet thick from (and resting upon) a hard rock like granite (Fig. 156).

Agents of Weathering. The atmosphere is the most important factor in the complex set of processes called weathering. The atmospheric gases and temperature changes are the outstanding agents. Additional help in the weathering processes comes from water and its dissolved chemicals and, to a minor extent, from lightning and organisms. The various agents may work together or may be combined in carrying on weathering. For example, water in rock crevices, through sufficient loss of heat and reduction of temperature, will freeze, expand, and disrupt the rock through the force of wedging. Such weathering requires two immediate agents, namely, water and temperature change.

Processes of Weathering. Broadly considered, there are two general groups of weathering processes—one mechanical, and the other chemical.

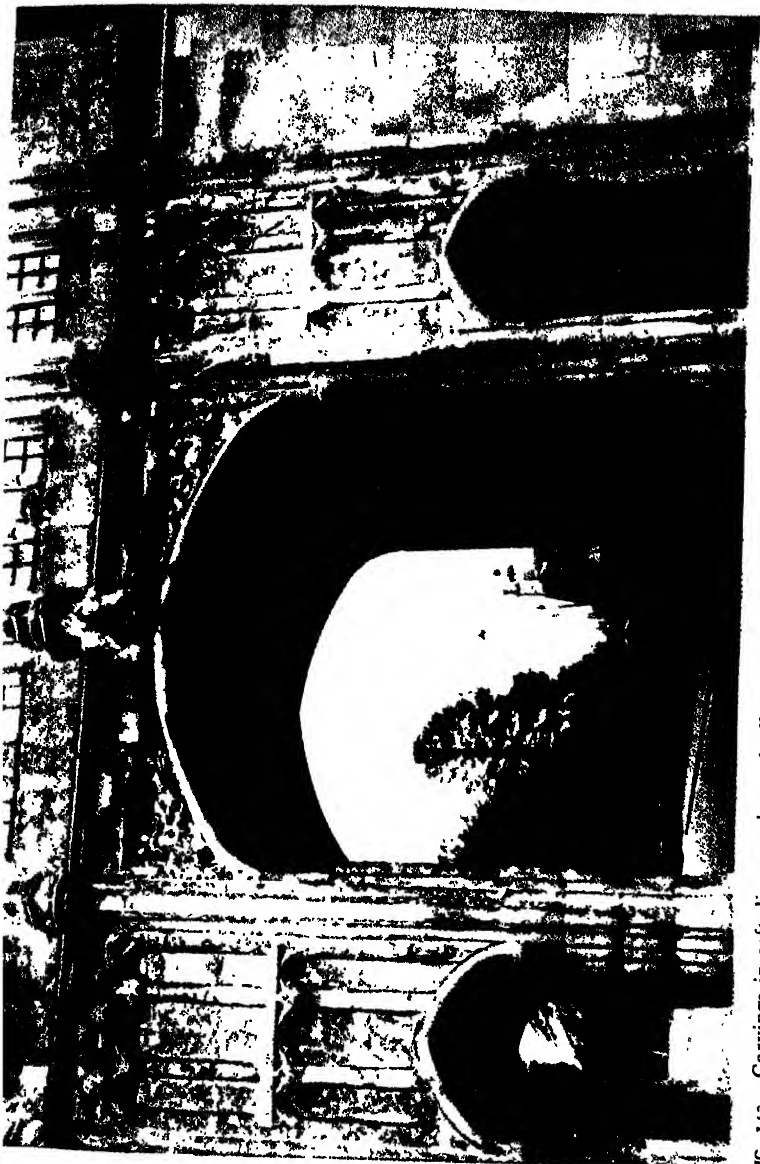


FIG. 142. Carvings in soft, limy sandstone badly weathered and disfigured by atmospheric agencies during some hundreds of years of exposure. Wells, England.

In *mechanical weathering* the rock breaks up or crumbles with little or no change in the composition of the material. It is essentially a physical process of disintegration. In *chemical weathering* the composition of the rock or mineral matter is more or less altered during its breaking up. It is essentially a process of decay or decomposition. Although the process of disintegration and decomposition may be thus distinguished, nevertheless the two processes very commonly operate together in nature, now one and now the other being predominant.

As stated by Blackwelder: "Rock weathering is due to the interplay of a number of processes. . . . Some of them are purely chemical, and others are strictly mechanical. . . . Doubtless some of these processes are everywhere of minor importance, while others may be powerful only at certain times or locally. . . . It is more than probable that one of these processes may facilitate a second, and that in turn a third. By such coöperation several agents of weathering may induce rock fracture where no one of them could do it alone."

Mechanical Weathering

Weathering by Freezing of Water. Not only in cold countries, but also in the mountains of regions with generally mild climates, alternate freezing and thawing of water constitute an effective agency in breaking up rocks, especially where soils are absent. Most relatively hard rocks contain not only numerous natural fractures called *joints* (see Chapter V) which separate them into more or less distinct blocks, but also small crevices, fissures, and pores. Surface water may fill such openings. Such water expands about one-tenth of its volume on freezing and exerts the tremendous pressure of many thousands of pounds per square inch upon the walls of the opening or fissure. If the rock is favorably situated, the pressure will widen the opening a little. Repeated freezing and thawing of water which finds its way into such openings finally causes even the hardest rocks to be mechanically broken up into smaller and smaller fragments. Jointed rocks situated on the faces of cliffs and steep slopes are especially subject to such action, as are also jointed or fissured boulders, pebbles, and even soil particles. This type of rock weathering is known as *frost action,* or, sometimes, as frost wedging.

Weathering by Temperature Changes Alone. For many years it was believed that effective disintegration of rocks in certain favorable climates resulted from the influence of temperature changes (*insolation*) alone. The theory was that all parts of a rock mass do not expand and con-

tract at equal rates when subjected to temperature changes, and therefore stresses are set up which cause the rock to break. Such effects were thought to be most conspicuous on high mountains and on deserts not only because rocks are there generally barren, but also because a daily range of 50 to over 100 degrees in temperature is frequent. In many deserts the outer portions of rock ledges and boulders exposed to the rays of the sun are heated to temperatures of 100° to 150° during the day and, therefore, expand, but during the night the temperature commonly falls 50° to 100°, and the outer portions of the same rocks contract notably. Cool rain of a sudden storm falling upon hot rocks also causes a quick lowering of temperature of their outer portions. Such rapid changes, causing repeated strains of this kind between the outer and inner portions of the rocks, theoretically should cause the rocks to break just as cold glass breaks when plunged into hot water. Most rocks consist of two or more kinds of minerals, each of which expands at a different rate; hence, additional, minor stresses and strains are set up, tending to pull apart the minerals and disrupt the rocks.

Many tests on rocks in laboratory experiments have failed to substantiate the theory, thus giving belief to the idea that in reality the disintegration effects on rocks in regions of high temperature ranges came about through slight chemical changes which may have been aided and abetted by temperature conditions rather than actually caused by them alone.

Blackwelder, who is among those who has questioned the importance of temperature changes as a factor in weathering, says "it is doubtful if insolation is ever the sole unaided cause of rock fracture. However, it seems at least possible that, by setting up strains in rocks, it aids in loosening the cohesion of minerals, facilitates the entrance of moisture, and thus promotes the breakage of the rock by expansion due to chemical changes." He believes that the principle is much the same as that involved in spheroidal weathering discussed later in this chapter. However, products of weathering, including feldspar grains, in desert regions are usually remarkably fresh, and so only minor chemical change as a factor of such weathering can be invoked. If the mineral grains of a rock become loosened by temperature changes, water can more readily enter the body of the rock so that, in sufficiently cold climates, freezing of the water will aid in disrupting the rock.

Exfoliation. Where, because of a combination of causes, usually including temperature changes, the outer portions of rocks peel, or scale, off in small or large slabs or sheets, the process is called *exfoliation*. It

seems to be especially effective in uniform, massive rocks, with fairly large mineral grains, such as granite or diorite. Rock surfaces tend to round off by this process, excellent examples being Stone Mountain in Georgia (Fig. 144), and many of the high mountains of the central and southern Sierra Nevada Range in California (Figs. 143 and 145). Not uncommonly, curved exfoliation slabs or sheets are hundreds of feet long and several feet thick.

There is reason to believe that, in many cases, large and small sheets of granite and other massive rocks shell off as a result of relief of pressure due to the removal by erosion of much overlying material (Fig. 83). This is explained as sheet jointing in Chapter V.

Mechanical Action of Organisms. Both directly and indirectly, plants and animals accomplish considerable work of rock disintegration. Roots and trunks of plants, especially of the higher forms, like trees, insert themselves in rock crevices and cracks, and as they grow they exert a powerful force, often sufficient to wedge the rock apart. Repetitions of this wedgework process often cause the rock to be broken into small fragments. Various rootless plants, such as lichens, attach themselves to rock surfaces and, by their growth, loosen off rock particles.

Indirect actions are the overturning of trees, causing relatively fresh rock materials to be brought to the surface and be better exposed to weathering agents, and the successive growths of roots, causing the soil to be made more open and accessible to weathering agents.

Burrowing animals, such as earthworms, ants, gophers, ground squirrels, and woodchucks, aid the action of weathering agents both by bringing fresher materials to the surface and by allowing more ready access of such agents to the surface materials. Earthworms perform a remarkable work of soil disintegration. They pass soil through their bodies in order to extract the vegetable matter from it, and in this way the bits of soil are ground up into still finer particles. It has been estimated that, in humid, temperate regions, the many thousands of earthworms per acre completely work over a soil layer from six inches to a foot thick once every half-century.

Other Mechanical Weathering Processes. Mechanical weathering is accomplished in some measure on relatively loose rocks and soils by the impact of raindrops and by the force of the wind whereby rock fragments are loosened from their positions. Lightning often shatters rocks in regions where electrical storms are frequent, but its total effect is relatively small.

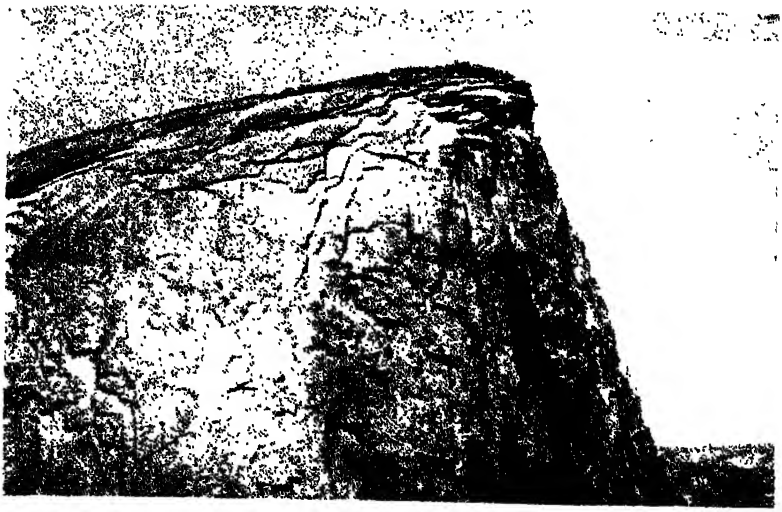


Fig. 143. Exfoliation of granite on a large scale. Half Dome, Yosemite National Park.

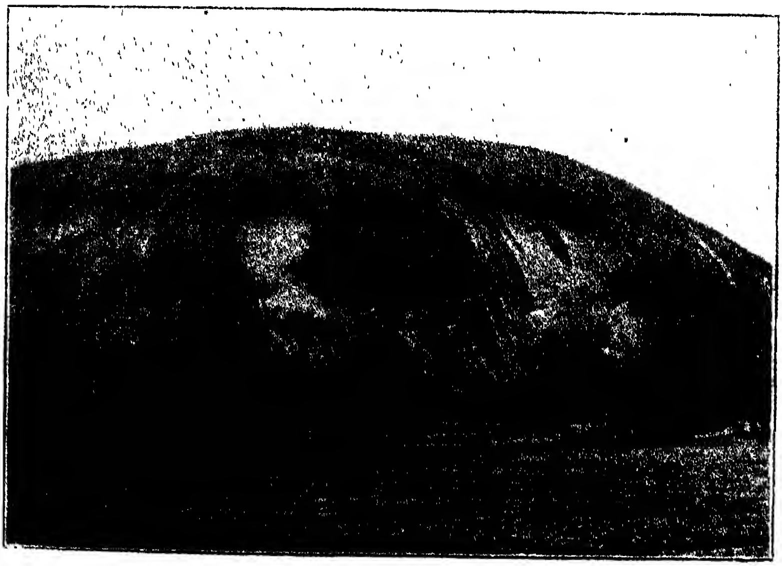


Fig. 144. Stone Mountain, Georgia, rounded off by exfoliation. (After Hillers, U. S. Geological Survey.)

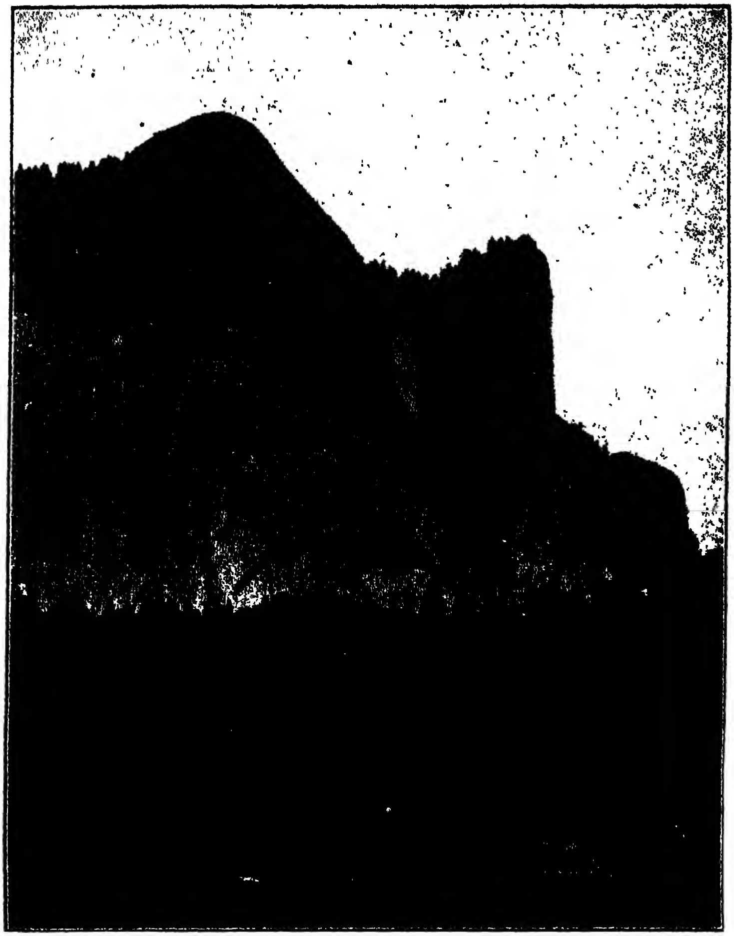


FIG. 145. The Royal Arches, Washington Column, and Basket Dome in Yosemite Valley, California, sculptured by exfoliation. (Courtesy of the U. S. Reclamation Service.)

CHEMICAL WEATHERING

Solution. Most rocks are only very slightly and slowly affected by the solvent action of perfectly pure water. Such water is, however, not found in nature because certain gases, particularly oxygen and carbon dioxide, are always dissolved in it, causing the solvent power of the

water to be notably increased as it becomes a weak solution of carbonic acid. Pure limestone is slowly but completely soluble in such water, and the dissolved matter is carried away by streams. In an impure limestone, only the impurities tend to remain as a residual soil. When rocks, whose mineral grains are cemented or held together by limy material, are subjected to the action of carbonated water, the limy material is dissolved and carried away, and the sand grains are left. Thus the rock crumbles. Many waters have their solvent power increased by the presence of other acids obtained from decomposing organic matter, volcanic gases, etc. If even only one kind of mineral in a rock is but slightly dissolved, the adhesion of the mineral grains is lessened, and the rock tends to crumble. Some minerals are exceedingly resistant to solution. Thus quartz is only slowly soluble even in hot, alkaline water. Such common minerals as gypsum and calcite are more or less readily soluble in hot, carbonated water. Salt is, of course, easily soluble in cold water.

Carbonation, Oxidation, and Hydration. Carbonic acid gas (carbon dioxide), which occurs in air, water, and soil, has the power of chemically uniting with, and altering the composition of, certain minerals of rocks. Thus many rocks contain the chemical elements calcium and iron with which carbonic acid gas may combine to form carbonates of calcium and iron. Such a process is called *carbonation*. The resulting carbonates, after conversion to the bicarbonate, are readily taken into solution and carried away by water, thus causing the rocks to crumble. A slow but very important and widespread process of this kind is the alteration of the very common mineral feldspar by carbonated water to kaolin (or clay), silica (or quartz), and a soluble carbonate. This action takes place during the decomposition of a hard, resistant igneous rock, like granite.

Oxygen occurs in air and soil, and dissolved in water. It is a very important chemical agent of decay of many rocks and minerals. The process of *oxidation* consists in the chemical union of oxygen with any chemical element, as very often happens with the iron contained in such common minerals as pyrite, biotite mica, hornblende (an amphibole), and augite (a pyroxene). The familiar rusting of iron involves oxidation, that is, a chemical union of the iron with oxygen of air or water.

The process of *hydration* consists in the chemical union of water with certain compounds. The principle is well illustrated by the rusting of iron which, on exposure to air and moisture, first unites with oxygen

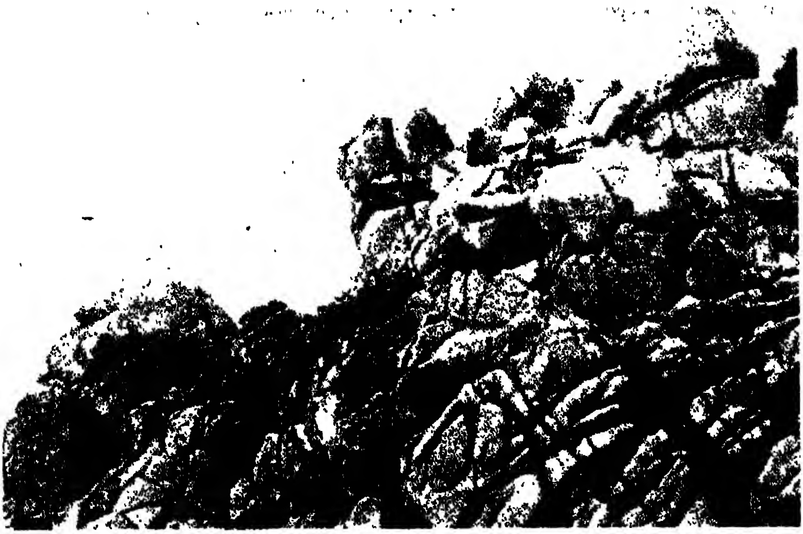


FIG. 146. An outcrop of granite showing how joints, being enlarged by atmospheric agents, facilitate both mechanical and chemical weathering of the rock. Near Lone Pine, California.

FIG. 147. An exposure of limy sandstone showing a cavernous appearance due to removal of certain more soluble and easily weathered material. Los Frijoles Canyon, New Mexico. (Photo by Mr. Harnden.)

to form iron oxide, and then unites with water to become yellow or brown hydrated iron oxide (the so-called "rust"). Many rocks contain iron not as such, but in chemical combination with other elements. When these iron-bearing minerals in rocks are subjected to the action of oxygen and water, the iron very commonly unites with both oxygen and water to form various hydrated iron oxides, ranging in color from yellow to reddish brown. Many of the striking colors of great rock

Fig. 148. A growing tree splitting a boulder of granite. Custer County, South Dakota. (Courtesy of the U. S. Forest Service.)

formations of the earth have thus been produced. An excellent, large-scale example of gorgeous coloring so produced is the Grand Canyon of Yellowstone Park whose iron-rich lava rock has been highly decomposed.

Carbonation, oxidation, and hydration are all very important factors in the chemical weathering, or decomposition, of rocks. Increase in volume of rocks affected is caused by all three processes, and the stresses and strains which develop as a result of volume increase tend to cause the rocks to crumble. In some cases the resulting materials, such as

carbonates, are dissolved and carried away by water, thus increasing the porosity and lessening the strength of the rocks affected.

Chemical Action of Organisms. Bacteria are very abundant not only in soils, but also on bare rocks. One group has the remarkable power of forming nitric acid from certain constituents (especially ammonia) of air, water, and soil, and this acid attacks and alters various minerals. Decaying plants, as well as roots of living plants, produce carbonic acid and other acids which alter the composition of various minerals in rocks.

Certain animals also bring about chemical weathering. Thus the soil particles which are worked over and carried by ants and earthworms are acted upon by organic chemical agents or acids secreted by these animals.

FIG. 149. Spheroidal weathering in basalt. Western Santa Monica Mountains, California. (Photo by U. S. Grant, IV.)

Spheroidal Weathering. When water containing dissolved gases enters a rock mass (particularly one which is fine-grained and homogeneous), which is divided into rectangular blocks by fissures, such as joint cracks (Fig. 82), the solutions work their way along the cracks and attack all surfaces of the rock with which they come into contact, and there cause decomposition which slowly eats into the blocks of solid rock. Not only do the edges, and still more so the corners, have

greater surfaces exposed to the solutions, but also they are attacked from two or three directions at once with likelihood of being affected by the strongest solutions. The corners of the blocks of rock will, therefore, most rapidly be weathered, the edges next most rapidly, and the faces least. The new substances thus formed by oxidation, hydration, and carbonation are greater in volume than the unaltered material, and so "strains are set up which tend to separate the bulkier new material from the core of unaltered rock. . . . The squared block is by this process transformed into a spheroidal core of still unaltered rock wrapped in layers of decomposed material, like the outer wrappings of an onion" (Hobbs). They are usually embedded in thoroughly decomposed material. The process described is called *spheroidal* or *concentric weathering,* and the resulting boulders are called *boulders of decomposition* (Fig. 149). It is to be noted that they are produced mainly by chemical weathering, whereas boulders of exfoliation result mainly from mechanical weathering.

Products of Weathering

Talus. A mass of rock fragments of various sizes and shapes resulting from the weathering of a cliff or steep slope and lying at the base

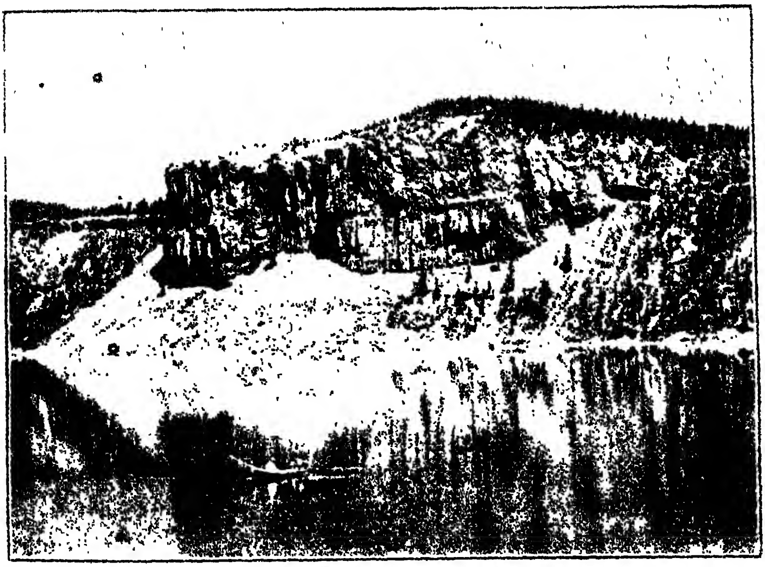


Fig. 150. A cliff of jointed lava and talus slope. Crater Lake, Oregon. (Photo by J. S. Diller, U. S. Geological Survey.)

of the cliff or slope is called *talus*. Exfoliation and freezing and thawing of water in cracks are the principal weathering processes which produce talus material. As the fragments are loosened from the cliff or steep ledge they fall, slide, or roll down until the angle of slope is too low for them to continue. The angle of slope of a talus pile generally ranges from about 25° to 40°. The tendency is for the largest blocks of rock to accumulate toward the bottom of a talus slope because the momentum carries such masses farther. In mountainous regions of

Fig. 151. Thick, horizontal beds of jointed sandstone undergoing weathering with resultant talus slope consisting of large and small blocks and fragments of the rock. Chatsworth, California.

severe climate with great and rapid changes in temperature, the conditions are especially favorable for large accumulations of talus, such deposits attaining lengths and depths of hundreds, or even thousands, of feet (Fig. 150).

Residual Boulder Fields. In many high mountains above the tree line and in the Arctic regions, great, barren, flat, or only moderately sloping rock surfaces are subjected to unusual rapidity of rock destruction mainly by frost action, that is, alternate freezing and thawing of water in cracks proceeds with such rapidity that the surfaces are often almost or completely buried under masses of shattered and broken-up

rock. Such rock fragments, which are generally angular in shape, may cover the bedrock from which they were derived to depths of 5 to 20 feet or more. Boulder fields of such origin are wonderfully displayed at and near the tops of both Long's Peak and Pike's Peak in Colorado at altitudes of 12,000 to over 14,000 feet (Fig. 152).

In many boulder fields caused by weathering, the boulders are more or less rounded (Fig. 154). Fields of such boulders may result pri-

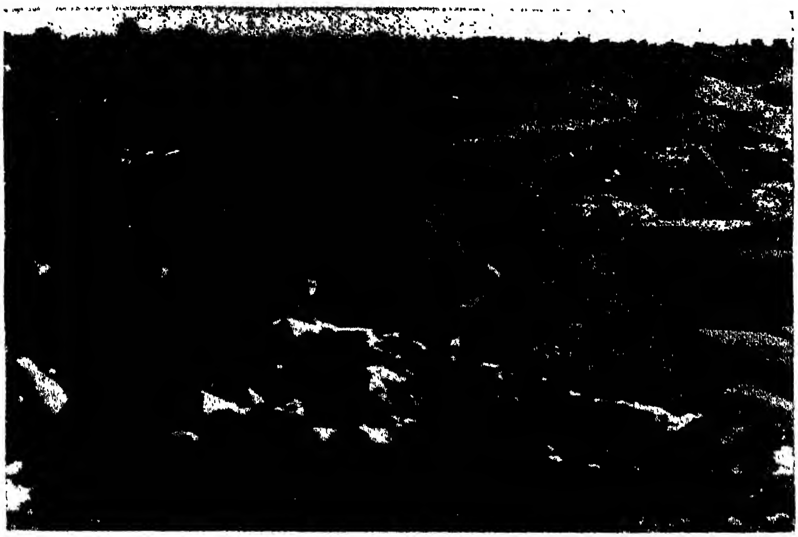


FIG. 152. Part of a field of angular boulders of granite resulting from mechanical weathering of the bedrock, mainly by frost action, at an altitude of more than 14,000 feet. Top of Pikes Peak, Colorado.

marily from mechanical weathering. Conditions are most favorable for their development in relatively dry regions not only because the process of exfoliation is there very effective in rounding off the original angular blocks of rocks, but also because the winds tend to keep the finer products of weathering blown away from between the boulders, thus keeping the bare surfaces of the latter constantly exposed to the weather.

Fields of rounded boulders may also result primarily from chemical weathering. Thus, the body of bedrock may be attacked much more vigorously by agents of decomposition along cracks, fissures, or more porous parts than it is in its more solid portions between the cracks or porous parts. There also may be local portions of the bedrock which are harder or more resistant to the weathering agents than the general

body of the rock. In either case, the tendency is for more or less rounded blocks of relatively fresh rock (Fig. 153) to be left as cores in the midst of highly decomposed material. Removal of the decomposed material by rain, streams, or wind will tend to leave an accumulation of the boulders at the surface. Some of the boulders in this category are boulders of decomposition already described as resulting from spheroidal weathering.

Mantle Rock. Most of the lands of the earth are covered by a superficial layer of loose, broken up material called *mantle rock* which,

FIG. 153. A road cut showing fresh boulders of weathering embedded in highly decomposed rock of the kind from which they were derived. An accumulation of such boulders covers the surface. The rock is diorite. Near Descanso, California.

wherever it occurs, rests upon the bedrock of the earth's crust. Where the bedrock is exposed at the earth's surface it is called an *outcrop*. There are two important kinds of mantle rock. One is the mantle rock which now rests upon the bedrock just where it was formed through the processes of weathering, that is, it is *residual mantle rock*, representing a direct accumulation of weathered rock material. The other is mantle rock which has been carried to its present position upon the bedrock, mainly by water, wind, or glaciers, that is, it is *transported mantle rock*, representing an indirect accumulation of material mostly made up of

FIG. 154. A highly jointed mass of diorite almost completely covered by boulders of weathering. Some of the jointed bedrock shows at the right. Seven miles west of Coyote Well, California.

FIG. 155. A boulder of weathering resting upon the ledge of granite from which it was derived. Another boulder is being weathered out on the left. Near Lone Pine, California.

products of weathering. Our present concern is chiefly with the residual rock mantle; transported mantle rock is treated in several of the succeeding chapters. The residual mantle does not rest by sharp contact upon the bedrock, but rather it grades downward through partly weathered rock into unweathered rock. The transported mantle rock rests characteristically by sharp contact upon the bedrock from which latter it usually differs notably in composition. Although the processes of weathering are universal and unceasing in their action over all the lands, nevertheless there are many places where conditions favor removal of the products of weathering fully as fast as they are formed, and so bare rock surfaces are left exposed.

The very widespread mantle rock is of great geological importance in several ways. Most of the vegetation of the land for countless ages has grown in its upper portion. It is the chief source of supply of the sediment carried by streams, and thus a great aid to erosion as we shall learn in a succeeding chapter. As an actual mantle it greatly retards the rapidity of weathering of the bedrock underneath it.

Soils. For the most part the soils of the world are either directly or indirectly the products of rock weathering. To some extent soils come from vegetation. In the strict sense of the word, *soil* is the relatively porous, fine-grained, upper portion of the mantle rock containing an admixture of vegetable matter and capable of supporting plant life. The term is, however, often used rather loosely. Just as we distinguish two general kinds of mantle rocks, so we must recognize two kinds of soils, namely, residual and transported. Residual soils here claim our chief attention because they are direct accumulations of products of weathering.

Residual soil, with its admixture of decomposing vegetable matter causing it to have more or less dark color, always grades downward into *subsoil* which usually contains fragments of partly decayed rock and little or no vegetable matter. The subsoil in turn passes by imperceptible change downward into partly decayed, so-called *rotten rock,* and this latter finally grades into the underlying, unaltered, so-called *fresh rock.* These various stages are well illustrated by Fig. 156.

Residual soils are very extensively developed in the southern states of the United States, and *transported soils,* left by the great glacier of the Ice Age, are very widespread over the northeastern states (Fig. 270). True soils are usually not more than a few feet thick, but soil plus subsoil and rotten rock may be scores or, exceptionally, hundreds of feet thick.

FIG. 156. Fresh granite (at bottom) grading upward through rotten rock and subsoil into true soil. Washington, D. C. (After G. P. Merrill, U. S. National Museum.)

Considering the large number of different minerals and rocks which give rise to soils and the varying conditions under which the materials are weathered, it is not surprising that there are many kinds of soils. In fact, probably no two soils from reasonably separate regions are just alike. Only a few of the more general soil types will be mentioned briefly. Thus, *clay* consists very largely of exceedingly finely divided kaolin. *Sand* is composed of sand grains, mostly quartz. *Loam* is a mixture of sand and clay. *Muck* is a very dark soil exceedingly rich in decayed vegetable matter. *Marl* is a soil rich in limy material, that is, in carbonate of lime. These very common kinds of soils show all sorts of gradations into each other.

Movements of Weathered Products

Attention has already been called to the accumulation of talus by the falling, rolling, and sliding of rock fragments which are loosened by weathering from cliffs and steep slopes. Closely related to this is the movement of rock débris in so-called *rock glaciers* or "stone rivers"

Fig. 157. A large landslide consisting of both weathered material and bedrock which broke away from the mountainside, moved rapidly down the slope, and dammed the river in the valley. Gros Ventre region, Wyoming. (Photo by E. Blackwelder.)

under certain conditions of cold climate. These are great masses of talus hundreds or even thousands of feet long, which slowly move down mountain sides or steep valleys, as in parts of Colorado. Externally they give somewhat of the appearance of a glacier. The motion results from

gravity aided by alternate freezing and thawing of water which fills the spaces between the rock fragments.

Soil creep is a common process by which mantle rock and soil move down slopes. When water-charged soil freezes, the rock fragments are lifted somewhat by the expansion at right angles to the slope or surface of the hill or mountain. On thawing, the rock fragments are pulled

FIG. 158. Unequal weathering and erosion of vertical beds of red sandstone. Garden of the Gods, Colorado. Parts of more resistant beds remain in bold relief.

down vertically by gravity, and thus they move downhill a little. Repetition of this process causes the whole soil mantle to move slowly or "creep" down the slope. There are also other causes of soil creep.

Rapid movements of masses of rock débris or bedrock down steep slopes are called *landslides* (Fig. 157). Among the causes of such movements are earthquake shocks; undercutting of the masses of débris by streams, thus weakening the support toward the bottom; and saturation of the mass with water, thus increasing its weight and lessening the friction of the rock fragments. In many cases not only the soil or mantle rock, but also much bedrock of a mountainside, takes part in a landslide. This happened at Frank in Alberta, Canada, in 1903, when

the whole face of a mountain several thousand feet high suddenly gave way, causing about 40,000,000 cubic yards of rock material to rush down into, and partly across, a valley (Fig. 175). Landslides are common, and many disastrous ones have occurred. Avalanches of snow also often carry much rock débris down with them.

Water, wind, and glaciers are by far the greatest agents of removal and transportation of mantle rock, including soil. The process of rock removal by such moving agents is called *erosion*. Water is most effective in humid regions; wind in arid regions; and glaciers in cold regions. All of these are very important geologically, and they are dealt with at some length in succeeding chapters. We shall here merely mention a few of the most important processes and effects involved.

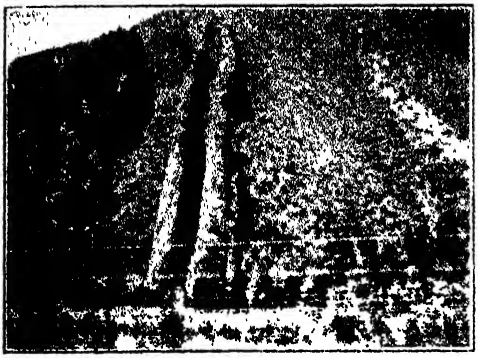


FIG. 159. Differential weathering and erosion of vertical strata. Two of the more resistant beds stand out. Devil's Slide, Webster Canyon, Utah.

By the direct action of rain wash, loose materials not too thoroughly protected by vegetation are carried from higher to lower levels. Streams carry tremendous amounts of sediment from higher to lower levels, the general destination being the sea. Much of the sediment is washed directly out of the mantle rock, but a considerable quantity is developed through the erosive action of the streams themselves as explained later. Much sediment is deposited temporarily on valley floors and becomes *alluvium,* especially on flood plains during floods. Some stream-carried sediment is deposited in lakes, and much of it forms delta plains and more widespread deposits in the sea at or near the mouths of the streams.

Large quantities of dust, soil, and small rock fragments are carried by wind. Two important types of transported mantle rocks originating in this way are dune sand and loess (see Chapter X).

Both valley glaciers and ice sheets (or continental glaciers) transport large quantities of rock débris. Thus, the vast glacier, which slowly moved over most of northern North America during the Ice Age, carried along and deposited so much rock débris that it is now by far the

FIG. 160. A hill of soft shale capped with a bed of hard sandstone. The shale slopes are strewn with angular blocks of sandstone derived from the well-jointed cap rock. Petrified Forest, Arizona.

FIG. 161. Volcanic tuff intricately weathered and eroded. Wheeler National Monument, Colorado. (Courtesy of the U. S. Forest Service.)

most common mantle rock and soil over the central-northern and northeastern states of the United States.

SCULPTURING BY WEATHERING

Effects of Differential Weathering. Most rock masses are not uniform in composition, texture, and structure. Some portions are, therefore, more readily attacked by agents of weathering than others, so that they are eaten into or etched out, while the more resistant parts are left

FIG. 162. A remarkably balanced rock resulting from unequal weathering of sandstone beds. Near La Veta, Colorado. (Courtesy of the U. S. Forest Service.)

to stand out in relief. All such cases of unequal weathering of rock masses are referred to as *differential weathering.* Such unequal weathering is a very common phenomenon to be observed in exposed bedrock almost anywhere. A few examples will suffice to make the principle clear.

Limestone or limy sandstone is particularly likely to become honeycombed, deeply pitted, or fluted where agents of weathering etch out the weaker and more soluble portions (Fig. 147).

Where a rock mass of any kind is transected by natural cracks called *joints,* the tendency is for the cracks to become enlarged while the intervening masses of rock stand out more and more separately in relief (Fig. 163). Where vertical joints cross-cut each other in closely spaced groups, leaving relatively large nonjointed blocks of rock between them,

FIG. 163. A row of joint columns, on top of a ridge, marking all that is left of a lava flow after differential weathering. Near Vantage, Washington.

the tendency often is for the weather to remove the jointed material and leave the solid cores which themselves become less angular under the action of the weather. Among many excellent examples are the Cathedral Spires in the Garden of the Gods, Colorado (Fig. 158), and the many wonderful natural monuments near Douglas, Arizona. Great joint blocks only partly etched out are wonderfully displayed in the walls of Zion Canyon, Utah (Fig. 79). A most remarkable maze of joint columns occurs in Bryce Canyon, Utah (Fig. 224).

Where veins or dikes of hard materials intersect ledges of weaker rocks, the vein or dike material often stands out in bold relief in the midst of the etched-out general body of weaker rock (Fig. 113).

Rock formations which are arranged in layers (usually stratified) often possess variable degrees of resistance to the weather. Where such rocks are in horizontal position or gently inclined, the tendency is for the more resistant layers to form cliffs or even overhanging ledges, while the weaker layers crumble down to talus slopes. Such differential

FIG. 164. Two almost completely separated cores of joint blocks marking all that is left of a once widespread, gently tilted sandstone formation. Both columns are differentially weathered. Garden of the Gods, Colorado. (Photo by N. H. Darton, U. S. Geological Survey.)

weathering is grandly displayed in the Grand Canyon of Arizona (Fig. 350). If the rock layers are steeply inclined or vertical, the tendency is for the more resistant layers to stand out in relief as sharply defined ridges (Figs. 159 and 343).

It is evident, from what has been said, that differential weathering plays an important part in the detailed sculpturing of the land. Many

of the more striking, minor features of landscapes, such as jagged peaks, pinnacles, ridges, and cliffs, have been so sculptured. Acting alone, however, differential weathering cannot proceed very far because the weathered products must be removed (eroded) by some agent such as wind or running water in order that new surfaces of the rock may be exposed to the weathering processes.

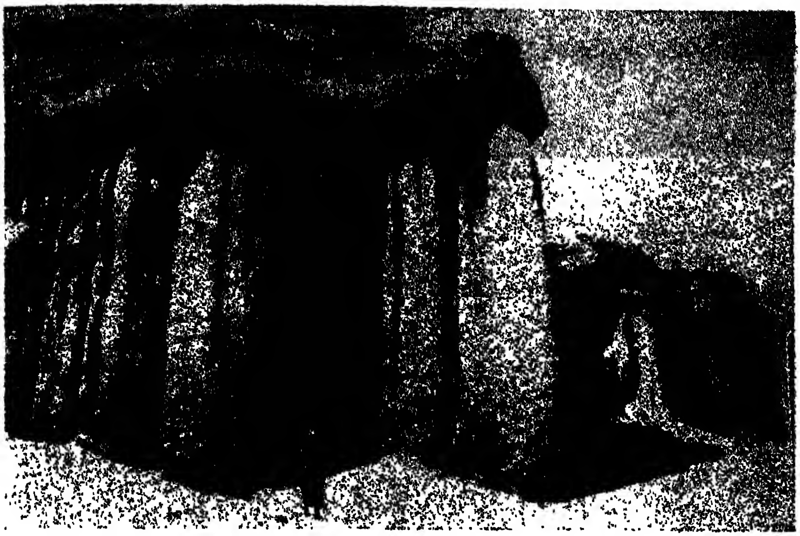


FIG. 165. Effects of differential weathering, aided by rain wash, in moderately consolidated, more or less bedded, volcanic, fragmental material. The fluted or columnar forms have developed in the weaker, vertically jointed portions of the outcrop. Red Rock Canyon, north of Mojave, California.

Effects of Exfoliation. The modes of origin of boulders by the mechanical weathering process of exfoliation and the chemical process of spheroidal weathering and also the manner of accumulation of such boulders into boulder fields have already been explained. There remains for brief consideration the more important topographic influence of exfoliation in the production of rounded or dome structures in rock ledges, hills, and even mountain peaks which may be called *exfoliation domes.* Hard, homogeneous rock masses, like granite with vertical joint cracks widely spaced, are, when exposed to rapid and great temperature range and other causes, favorable for development of large-scale exfoliation domes because the rock scales off in large slabs up to several feet thick and scores or hundreds of feet wide and long. As the

successive slabs peel off, the rock masses gradually become more curved or convex, thus giving rise to curved surfaces and dome structures. Excellent examples are Stone Mountain, Georgia (Fig. 144), and various mountains of Yosemite Park, California (Fig. 145). Where several sets of joints are closely spaced and roughly at right angles to each other, the rock generally breaks up (under the weather) first into rectangular blocks resembling crude masonry (Fig. 82) and finally into a mass of boulders. Where only vertical joints are well developed, the rock tends to break up into jagged cliffs, ridges, and needlelike summits (Fig. 224).

CHAPTER VIII

THE WORK OF STREAMS

Introduction

Importance of Streams. All things considered, running water is the most important of the three great agents of erosion—water, wind, and ice. About 30,000 cubic miles of water (partly in the form of snow) fall yearly upon the lands of the earth. Approximately one-fifth of this tremendous quantity of water is carried by streams into the sea each year. Some idea of the enormous amount of energy developed by these streams may be gained by a statement of the fact that they make an average decent of nearly one-half of a mile, this being the average altitude of the lands of the earth. A very considerable amount of energy is also developed by streams in desert regions which do not enter the sea. Much of this energy is used up in friction, in wearing away rock materials, and in transporting sediment.

FIG. 166. Detail view showing effects of rain wash upon slightly consolidated sediments. Near Laguna Beach, California.

Sources of Stream Water. Rainfall and snowfall constitute the proximate source of almost all stream water. A large part of the water of most streams results from the immediate run-off of the rainfall. The amount of such water fluctuates greatly. The melting of snow, especially in the spring in the northern hemisphere, contributes much water to streams. Another source of supply is glaciers, nearly every one of which has a stream emerging from it. Ponds and lakes commonly feed water into streams. Still another source of stream water is subsurface

water which emerges at the surface in the form of springs. This supply tends to fluctuate relatively little, and it is, therefore, an important factor in the regulation and maintenance of stream flow, especially during periods of dry weather. A large river like the Mississippi may receive important contributions from all the sources above mentioned.

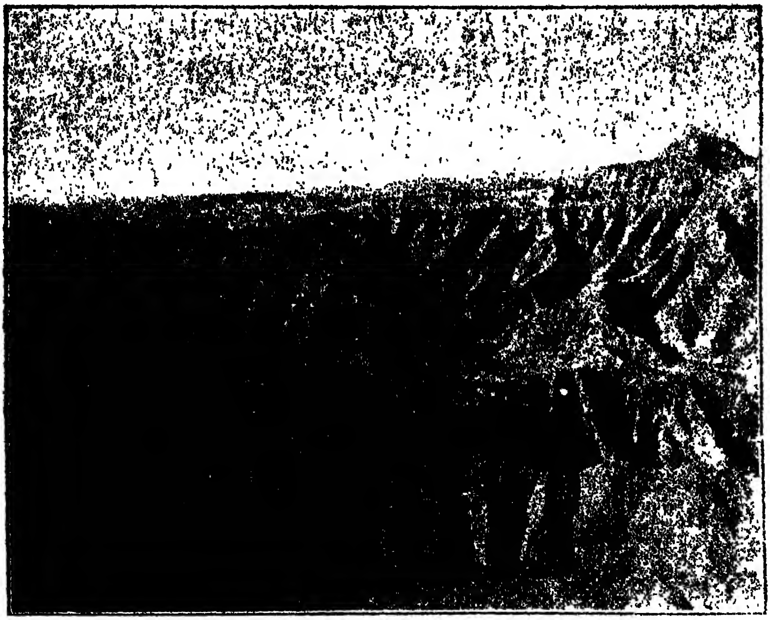


FIG. 167. A view in the badlands of South Dakota showing rain wash effects and gully development in soft strata. (After Barnett, U. S. Geological Survey.)

Rain Wash. Rain water accomplishes a certain amount of erosion before it collects into definite streams. Everyone is familiar with the fact that soils are carried down slopes by the wash of the rain. When rain "flows off in a sheet, as on a smooth surface, the depth of the water is slight, the flow not very swift (unless the slope is very steep), and the wear correspondingly slight. Such wear is called *sheet erosion*" (Chamberlin and Salisbury).

When rain falls upon soft or loose material such as clay or sand, especially where the slopes are steep, the effect of rain wash is much more pronounced than when it falls upon hard rocks (Fig. 166). A mantle of vegetation, of course, tends to protect soils and rocks against

FIG. 168. A large outcrop showing effects of rain wash and weathering (including solution) in nearly horizontal beds of moderately consolidated, vertically jointed, limy strata. Bryce Canyon, Utah. (Photo furnished by Union Pacific Railroad.)

rain wash. For these reasons it is easy to understand why, in regions like New England and the Adirondack Mountains, where bedrock formations are very hard, and where soils are mostly stony and resistant, the streams are generally clear except in times of flood, while in regions like the *badlands* of South Dakota or parts of the western Great Plains or even parts of the southern states, where not only the soils but also the bedrock formations are soft, the streams are very commonly muddy or turbid.

Methods of Stream Erosion

The place of *erosion* among the several processes of gradation has already been indicated in the discussion of rock weathering. Erosion is a rather complicated process which may be defined simply as the natural removal and transportation of rock material. It comprises all processes whereby the lands are worn down. It is carried on by moving (dynamic) agents, most significant of which are running water, wind, sea, and moving ice. The actual methods (subprocesses) whereby erosion is carried on differs among the various agents. Stream erosion involves the following subprocesses: (a) *corrasion,* (b) *solution,* or *corrosion,* (c) *hydraulic action,* and (d) *transportation. Transportation,* as a separate, important process of erosion, is reserved for fuller discussion later. Unless rock materials which enter streams are carried away (transported), there can be no wearing down of the land.

Corrasion. Surfaces of hard rocks are only very slightly worn mechanically by clear water flowing over them. A remarkable case in point is the constant rush of the mighty volume of clear water over the crest of Niagara Falls. The water is clear because it comes out of Lake Erie which acts as a settling basin for sediment. In spite of the velocity of the water, myriads of tiny, growing plants (algae) are attached to the rocks at the very brink of the falls, proving that any mechanical wearing away of the rocks must be very slight. The same principle is illustrated by many very clear streams with even rather swift currents which emerge from lakes and whose sides and bottoms may be lined with moss or other vegetation.

In cases of streams of at least moderate velocity and properly supplied with grinding tools mechanical erosion becomes very pronounced. The tools are rock fragments of all sizes from those of silt, mud, or sand to pebbles or even boulders. By *corrasion* is meant mechanical wearing away of rocks by the rubbing, grinding, and bumping action of rock

fragments carried by any agent of transportation against the bottom and sides of the channel and also against themselves.[1] It includes abrasion.

From the above statements it may be readily understood that the factors which facilitate rapid corrasion by streams include swift current, a liberal supply of tools (especially of angular fragments of hard rocks and minerals), and relatively soft or weak rock over which the water

FIG. 169. Tools with which a river works. Boulders in the bed of a river at time of low water. Near Wells, New York.

flows. Since the tools themselves are worn in the process of corrasion they soon become rounded. This is true of all sizes of rock fragments from the tiniest to big boulders.

Solution. We have already learned that many rocks and minerals are more or less soluble in water, and that their solubility is increased by the presence of small amounts of carbonic acid gas and oxygen which are found in all water in nature. Limestone is, of all the very common rocks, most susceptible to solution, being in fact completely soluble when perfectly pure. Although the process is a slow one as measured

[1] As used in this book, the term *corrosion* is broader in its scope than *abrasion*, which includes only rubbing or grinding action. The term *corrosion* implies solution or chemical action.

by the span of an ordinary human life, nevertheless a stream of even moderate velocity flowing over bedrock of limestone or even impure limestone carries away a large amount of the rock in solution in a short time, geologically considered. Such a process carried on by running water is frequently referred to as *corrosion*. Where a stream flows over a hard igneous rock like granite, the work of solution is very much less effective because the quartz in this rock is scarcely, if at all, affected, and some of the feldspar material goes into solution only very slowly.

FIG. 170. Jointed bedrock being little affected by a stream at time of low water. At times of high, sediment-laden water both corrasion and plucking are effective over the whole surface. Adirondack Mountains, New York.

As a result of rain wash over the ground and on gully or valley sides, more or less mineral matter is taken into solution and carried into streams. A large river like the Mississippi carries a tremendous amount of dissolved mineral matter into the sea each year. This phase of the subject is treated a little beyond in this chapter.

Hydraulic Action. The mere impact or pressure of running water may, under certain conditions, effect a considerable amount of erosion, even without the aid of tools. Thus a stream of relatively clear water, flowing through soft or loose material, may by this process cut back its

bank or push off material from the bottom of the channel. But even where rocks are hard they are very commonly intersected by numerous cracks (so-called *joints*), causing the rocks to be more or less separated into angular blocks. In sedimentary rocks the stratification surfaces are often also a factor in dividing rock masses into blocks. In many places such joint blocks are only loosely attached to the parent ledge, especially where various agencies of weathering have acted along the joint cracks. Many loosely attached joint blocks of this kind are removed by

Fig. 171. A detail view of the runaway Colorado River cutting into banks of soft, deep loamy soil as it rushed through the Imperial Valley, California, between 1904 and 1907. (Photo by U. S. Geological Survey.)

the mere pressure of the current flowing against them. The subprocess, or method, of erosion described above is commonly referred to as *hydraulic action*. The agent is the pressure of running water.

Influence of Joints in Stream Erosion. Mention has already been made of the influence of joints in aiding streams to push off blocks of rocks from ledges by mere impact ot the current. Where running water enters joint cracks in the sides or bottom of a channel, the work of solution is increased because much larger surfaces of rock are exposed to the action. In limestone or limy rocks, joint cracks are often so enlarged by solution that the joint blocks become easy prey to the pressure of the

current which pushes them away. The work of corrasion is also made more effective by joints, particularly where they have been enlarged by solution or by other weathering agencies, because more rock surfaces are then exposed to corrasive action along the bottom and sides of the channel.

Transportation by Streams

Nature of the Stream Load. All the material carried by a stream constitutes its *load.* The visible load consists of materials carried in

Fig. 172. Channel cut by flood waters in about thirty minutes, during 1923. Mouth of Willard Canyon, near Ogden, Utah. (Photo by Robert Nevins.)

suspension and rolled or pushed on the bottom, whereas the invisible load is the mineral matter carried in solution. The first is a mechanical load; the second, chemical. What are some of the sources of the visible stream load? This question may be readily answered by following a typical, small, swift stream, especially in time of flood, through its valley for a few miles or even less. Materials are carried down the valley sides or slopes by rain wash or by water from melting snow; landslides and avalanches, as well as the slower movement of hillside creep, contribute considerable quantities of rock fragments of all sizes; loose deposits of clay, sand, gravel, and even boulders, through which the channel is being cut, easily become part of the load; solid rock of the valley walls, where undercut by the current, falls into the stream; joint blocks of both bottom and sides of the channel may be pushed off by

the pressure of the current and become part of the load; and many fragments are supplied by the process of corrasion on the sides and bottom of the channel.

Much of the material in solution (invisible load) is supplied to streams by underground waters where they emerge as springs in the stream valley; some is added as a direct result of the solvent action of rain water on the valley sides; and some is taken into solution by the stream itself from the rocks over which it flows.

The water in a stream does not move as a simple forward current, but rather it is subjected to very complicated motions including the main current forward, upward, and downward movements, and "eddies" and "boils." The secondary upward currents bring the finer rock fragments (sediment) into suspension by carrying them up from the bottom of the stream. The coarser, heavier rock fragments are either pushed or rolled along on the bottom of the stream and constitute the bed load.

Influence of Velocity. Even a moderate increase in the velocity of a stream increases almost incredibly its power to transport rock débris. In addition to velocity of current, the transporting power is variably conditioned by size, shape, and specific gravity of the rock fragments carried, volume of water, and shape of the channel. A stream becomes swifter not only with increased gradient or declivity of the stream bed, but also with increased volume of water. The statement made by various writers in the past that the maximum size of rock fragments carried as bed load by a stream varies as the sixth power of the velocity is largely or wholly theoretical. According to that statement, doubling the velocity of a stream could increase its transporting power as much as 64 times. More recent studies and experiments, however, indicate that such a maximum is rarely if ever attained in nature because of the variability of the factors involved. In any case a mere doubling of velocity certainly increases the carrying power of a stream manyfold, possibly as much as 32 times.

When a stream rises very rapidly during a cloudburst, or when a dam suddenly gives way, the water rushes down a valley with high velocity. An understanding of the effect of increased velocity on stream transportation helps us to comprehend why, under such circumstances, the water does so much damage, carrying along massive bridges, big boulders, and even locomotives, as happened during the famous Johnstown flood of 1889. In 1928, when the St. Francis dam gave way in Los Angeles County, California, a block of concrete weighing about

10,000 tons was carried downstream about one-third of a mile. It is obvious, therefore, that even streams which are swift at times of low water have their power to transport rock débris greatly increased in times of floods. Not uncommonly a stream will, in a few days of flood, transport more material than in many days or even weeks of low water.

It has been proved that a current with a velocity of only six inches per second can carry along fine sand; one flowing one foot per second (or about two-thirds of a mile per hour) can move ordinary gravel; four feet per second (or nearly three miles per hour) pebbles weighing

FIG. 173. Effect of flood water which started on the south face of Mount Hood, Oregon, in 1930. The whole boulder-strewn surface here, and for miles up and down stream, was covered by a raging torrent which transported the countless boulders together with a great amount of finer material. Mount Hood is in the background.

about two pounds; and 30 feet per second (or 20 miles per hour) boulders weighing hundreds of tons.

In all of our considerations of stream transportation, it should of course be borne in mind that, due to the buoyant action of water, a mass of average rock with a specific gravity of nearly three loses about one-third of its weight when immersed in water. This greatly facilitates the transporting power of currents.

Graded and Overloaded Streams. Most streams have sufficient velocity and volume to transport more material than is fed into them

from tributary slopes and streams. Such a stream, therefore, has energy left to cut down its channel, that is, to *degrade* it. As the down-cutting process goes on, the *gradient* (or declivity) of the stream bed becomes more and more gentle until a condition is reached in which the whole energy of the stream is used up in transportation, and then degradation ceases.

Some streams are unable to transport all the material which is fed into them. They are said to be *overloaded.* Not only does such a stream lack power to cut down its channel, but it actually deposits part of its load and so builds up its channel, that is, *aggrades* it. In certain cases, where such aggradation goes on, the gradient of part of a stream may gradually become steeper until the stream is there able to transport its whole load.

A stream which has reached the balanced condition between down-cutting and deposition is said to be at grade. In other words, a *graded* river is one which, on the average, neither degrades nor aggrades, but is just able to carry the load supplied to it from tributary slopes and streams. Owing to varying conditions, portions only of a stream may be temporarily graded. Also, a graded stream may degrade its channel during times of flood, while during times of lower water it may deposit, but it is the average condition which should be considered.

Amount of Material Transported. Within the lifetime of a human being, the ordinary river seems to accomplish little or nothing by way of enlarging its valley. Within a relatively short part of geologic time, however, a large valley or even a great canyon may be carved out (eroded) by a stream. Thus, what is now the space occupied by the whole Connecticut Valley of western New England was once filled by a mass of solid rock which, during the present (Cenozoic) era of geologic time, has been weathered and eroded, and the resulting materials carried away by the Connecticut River. Or again, the mighty Grand Canyon of Arizona has been formed since the middle of the present era as a result of the removal of a body of rock hundreds of miles long, 8 to 15 miles wide, and thousands of feet thick (Fig. 217). So it is with nearly all the valleys and canyons of the world because, with relatively few exceptions, they have been carved out by the eroding and transporting power of the streams which they contain. It is, as would be expected, a general rule that the larger valleys are occupied by the larger streams.

Fig. 174 illustrates profound erosion in another way. The strata

flanking the granite and schist on each side of the Front Range of northern Colorado formerly extended over the whole range in the form of an anticlinal blanket about two and one-half miles thick and 50 miles across. During and since the formation of the anticline, at the beginning of the present (Cenozoic) era of geologic time, this great sedimentary cover has been carried away by streams.

According to a good estimate, about 1,000,000,000 tons of material in suspension, solution, and rolled along are carried annually by the rivers of the United States into the sea. Some conception of the amount of this material may be gained from the fact that a train of ordinary

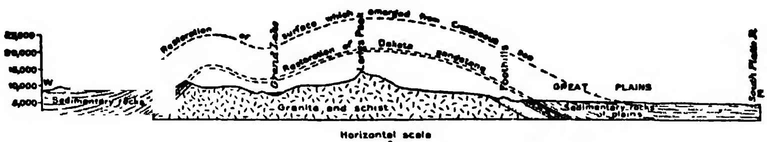


FIG. 174. Structure section through the Front Range of northern Colorado showing how a great anticlinal blanket of strata has been removed by erosion. (After W. T. Lee, U. S. Geological Survey.)

freight cars long enough to contain it would reach around the earth at the equator more than six times!

The Mississippi River drains more than one-third of the area of the United States proper. Older observations and tests indicated that this great river discharges annually nearly 600,000,000 tons of material into the Gulf of Mexico. More recently R. J. Russell has concluded that the annual discharge is nearer 700,000,000 tons. The difference is due mainly to uncertainty in regard to the amount of sediment rolled or dragged on the river bottom (bed load). More than one-half of the stream load is material in suspension, about one-fourth is material in solution, and the rest is bed load. All the material carried represents rock material removed from the Mississippi River drainage basin which covers a million and a quarter square miles. It should be remarked that much of the material in solution is supplied to the river by springs, the waters of which dissolve the material during their underground travels, mostly within a few hundred feet of the surface as explained in Chapter XII. In view of the facts that the Mississippi basin is so vast, and that it includes such a great variety of topography, rocks, and climate, the river which drains it is, in proportion to its size, about an average one as regards the amount of burden carried.

Rate of Stream Erosion

Rate of Erosion of the Mississippi Basin. All lands are being more or less cut down (eroded) by streams, and estimates of the rate at which certain rivers are lowering their drainage basins have been made. As a result of measurements and tests near the mouths of these rivers, the load of material carried yearly in suspension, solution, and rolled along by each of them has been determined. Since the burden represents rock material which has been removed from the whole drainage basin of a given stream, and the area of the basin is known, it is easy to calculate how thick a layer of this material of uniform depth would be if spread over the whole basin. The result, of course, represents the average yearly rate at which the drainage basin is being eroded. Data regarding the Mississippi River are unusually accurate. The area of its drainage basin is 1,265,000 square miles. As a result only of the material carried in suspension and solution, the Mississippi Basin is being lowered at an average rate of one foot in approximately 6120 years (U.S.G.S. Water-Supply Paper 234). Considering also the amount of material rolled along the bottom of the river, the drainage basin of the mighty river is being lowered at an average rate of one foot in about 5000 to 5500 years. Although it should be understood that this figure is the result of only an estimate, it is nevertheless probably accurate to within 10 per cent, and thus gives a good idea of the order of magnitude of the rate of erosion by the Mississippi River. In regard to rate of stream erosion, the Mississippi is probably not far from the average for the streams of the world which enter the sea. Some, however, erode much faster, and others much slower. Thus it has been estimated that the Ganges River of India cuts down its drainage basin more than twice as fast as the Mississippi. This is not only because of the much greater average swiftness of the river, but also because of the unusually heavy rainfall over its basin. In the case of the Danube River of Europe the rate of erosion is much less, being one foot in nearly 7000 years.

Rate of Erosion of the United States. According to estimates made by the Geological Survey, the rivers of the United States transport yearly to tide water nearly 800,000,000 tons of material in suspension and in solution, but this figure is probably too low. Adding the less accurately known amount of bed load material, the grand total transported to tide water yearly by the rivers of the United States is probably about 1,000,000,000 tons. If the erosive energy thus expended could

have been concentrated upon the Isthmus of Panama, the work of excavation for the canal would have been accomplished in less than two months.

All factors considered, and using the best available estimates for the rivers, the surface of the United States is being worn down at the rate of one foot in about 8000 years. This is slower than the rate for the Mississippi River because the large area called the Great Basin in the western interior of the country has no drainage outlet whatever to the sea, and other areas like Florida are so low that the average level of the land is either not being lowered or the rate of erosion is notably less than for the Mississippi Basin.

Time Necessary to Wear Away North America. The rate at which the continent of North America is being cut down is probably approximately the same as that for the United States, that is, one foot in about 8000 years. Since the average height of North America is about 2000 feet it is clear that the streams eroding at their present rate could not cut it down to sea level in less than 16,000,000 years. As a matter of fact, the time required would be much longer than 16,000,000 years, for, as the land would gradually become lower, the power of the streams to erode, or, in other words, the rate of erosion, would steadily become less and less. When we realize that large and small portions of continents have been worn down to the condition of plains, or nearly so, during various periods of the known history of the earth, we are forced to conclude that running water has been at work on the face of the earth for many millions of years.

In this general connection the reader should bear in mind that whole continents would seldom if ever be even approximately leveled, because, within such wide areas, constructive forces of diastrophism or volcanic activity would operate to maintain the land.

Valley Development by Streams

Most Valleys Formed by Stream Erosion. Nearly all streams flow in more or less well-defined valleys. Most of these streams by far flow through valleys which have been carved out by the erosive work of the streams. Some reasons for so believing are that valleys vary in size according to the size of the streams which occupy them, that is, the larger the valley the larger the stream in it; tributary streams and valleys are smaller than the ones they join; a vast majority of tributary valleys and their streams enter larger valleys at accordant levels, that

is, at the same elevation as the floors of the larger valleys; and many streams, aided by ordinary processes of weathering, are definitely known to be deepening and widening their valleys.

Some valleys were, however, ready-made for the streams which occupy them. These are usually *structural valleys,* so-called because they have been formed by earth-crust movements (*diastrophism*). Some structural valleys, like the Owens Valley in California and the Jordan Valley in Palestine, each many miles long and thousands of feet deep, were formed by the subsidence (down-faulting) of long, relatively narrow blocks of the earth's crust. Certain others, like the Great Valley of California, were formed by uplift of land into hills or mountains on either side of the valley.

Beginning of Valleys. A new land surface formed in any manner, as for example by the draining of a lake or by the uplift of land (out of the sea in many cases), soon has a drainage system established upon it. Water from rainfall or melting snow does not flow uniformly over the more or less uneven new surface, but it very soon tends to concentrate in the depressions and begins to run off in streams. These initial streams begin to carve out *gullies* which, with every fresh supply of water, become deeper, longer, and wider (Fig. 167). After a time the gullies are large enough to be called *valleys.* Many gullies may start on a new steep slope, but as times goes on certain of them become wider and take in smaller adjacent ones, and so relatively few of the original gullies really ever become valleys of considerable size.

Not all stream-cut valleys have started their development in the form of gullies. Thus over a great portion of northern North America the vast glacier of the Ice Age (Chapter IX) left widespread, irregular accumulations of rock débris over large portions of the area from which it retreated by melting. "Large parts of the surface were left without well-defined valleys, but with numerous lakes (e.g., Wisconsin and Minnesota). The rainfall of the region was enough to make these lakes overflow. Where a lake overflows, the outgoing water follows the lowest line accessible to it, so long as there is a line of descent. In this case, the running water will start to cut a valley all the way from the lake which furnishes the water to the end of the stream, at the same time. No part of such a valley is much older than another." (R. D. Salisbury.) Such a valley is, of course, not a grown-up gully.

Valley Lengthening. Water which flows into the upper end of a gully or valley cuts back its head by erosion. By such a process of

headward erosion, a gully or valley is lengthened. A valley head is seldom cut back (lengthened) in a straight line. One reason for this is that differences in the character of the rock material cause headward erosion to proceed more easily in some places than in others. Another is that irregularities of the surface cause the water which flows into the head of the gully or valley to be concentrated first in one position and then in another. For such reasons the headward erosion proceeds irregularly, and thus the crookedness of so many valleys is accounted for.

If a large spring is located at the head of a valley, the dissolving and undermining action of the spring water may aid headward erosion considerably by recession of the spring head.

The lengthening of a valley ends when a permanent *divide* (division of drainage) is developed, because then the amount of erosion on one side of the divide is counterbalanced by that on the other side.

Valley Deepening and Base Level. A valley is deepened by the cutting down of its floor by the erosive action of the stream which flows through it. The excavation of the valley is brought about by all the processes of erosion—weathering, corrasion, solution, pressure, and transportation—which we have already discussed. According to varying conditions, such as swiftness and volume of water, supply of tools, climate, rock character, and rock structure, these different erosive processes operate with varying degrees of effectiveness.

As time goes on, and if no other process intervenes, the power of a stream to cut down (degrade) its valley gradually diminishes, because of lessening velocity, until a limit is reached below which the stream cannot cut. The limit is the level of the sea, lake, or valley floor into which the stream empties, but obviously only the lower course of the valley can ever actually reach the limit because there must be at least a slight slope (*gradient*) farther up the valley in order that the stream may continue to flow. The lowest level to which running water can cut down (degrade) a land surface is called *base level*. In this connection it should be remarked that the channel of a stream may be actually a little below the level of the standing water into which it flows. Thus at and near the mouth of the Mississippi the *channel* of the river is as much as 100 feet below tide water because the current of the mighty river is able to keep its channel scoured to that depth as it rushes into the Gulf of Mexico.

It follows as a corollary to the preceding statements that very deep stream-cut valleys and the special forms of valleys called canyons can

be developed only in lands high above sea level because the higher the land the deeper can a stream erode before approaching base level. This explains why very deep valleys and canyons are found invariably in plateaus and mountains, as for example the Grand Canyon of Arizona in the Colorado Plateau, and the very deep, narrow canyons, such as Yosemite and Kings River, in the Sierra Nevada Range of California.

Valley Widening. Most valleys are much wider than the streams which flow through them, but it by no means follows that the streams were ever necessarily wider or larger than they now are. If a valley developed wholly by the down-cutting action of a stream, the valley would be no wider than the stream, and its walls would be vertical. This latter type of valley (or rather gorge) is approached where all conditions for down-cutting are so favorable that they greatly predominate over other factors which operate to widen the valley. An excellent case in point is the upstream youthful portion of Zion Canyon, Utah, which is over 1500 feet deep with nearly vertical walls which are in places not more than a few hundred feet apart at the top.

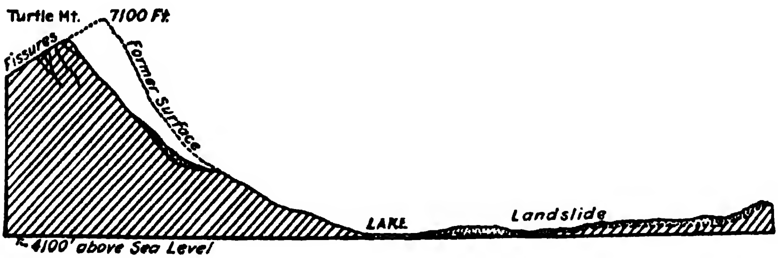


FIG. 175. Structure section illustrating the great landslide at Frank, Alberta, Canada, in 1903. (After McConnell and Brock.)

Some of the ways by which the great majority of valleys are made wider across their tops are the following: Loose, weathered materials are washed down the valley sides by rain. If the slopes are steep, some materials roll down, and loose materials, especially when they are soaked with water, may *creep* or *slump* to lower levels. On steep slopes, rock material may move down suddenly in the form of *landslides* (Fig. 175). Talus piles accumulate at the bases of very steep valley walls as a result of weathering. Materials which move to the bottoms of valley sides in these and other ways are usually carried away by the streams in the valleys, and thus the tops and sides of the valleys which are occupied by

actively down-cutting streams steadily become wider. Then, too, since streams are rarely if ever straight, the current in many places strikes one side of the channel with greater force than the other. Thus, while a stream is cutting its channel deeper, it is also doing some direct work of lateral erosion and so widening its valley at the bottom. Valley widening of this kind is, however, mostly accomplished by streams at or near grade as will next be explained.

Lateral Erosion and Its Results. Some work of lateral erosion is accomplished by rather actively down-cutting streams, as just explained,

Fig. 176. A meandering stream in a small valley. South Russell, St. Lawrence County, New York.

but the most effective work of this kind is done by streams with relatively low velocities and little or no down-cutting power, that is, by streams at or near a graded condition. Such a stream may flow upon an original nearly flat surface, or in any valley, or portion of a valley, where a graded or nearly graded condition has been reached. In a slow-moving stream of this kind, the current is easily turned against one side or the other of its channel. This may be brought about where the swifter current of a tributary enters, or by some obstacle like a rock, or where the material of one bank is more easily eroded than that on the opposite side.

If for any reason the main current of a slow-moving stream strikes with greater force against one bank, it will be eroded sidewise, and from there the current will be deflected against the opposite bank somewhat farther down stream, causing lateral erosion to take place there. Such side-wise cutting is frequently referred to as *lateral planation.* By a continuation of such a process, with the points of attack shifting downstream little by little, a series of sweeping curves called *meanders* develops (Fig. 176). Such meanders become more and more pro-

FIG. 177. **The oxbow of the Connecticut River near Northampton, Massachusetts. The meander was cut off in 1841.**

nounced as a graded condition is approached by the stream, and they finally become a series of loops mostly separated by only narrow necks. Finally the necks are cut through one by one, and cut-off meanders, called *oxbows,* are formed, marking the old channel. Meanwhile other meanders and loops develop.

Wide flats, called *flood plains* (Fig. 184) because they are flooded at times of high waters, are developed by this process. The lower reaches of some great rivers, as for example the lower Mississippi River, have developed flood plains 20 to 75 miles wide and hundreds of miles long. Farther and farther upstream the flood plains usually become less and less prominent. Meanders and oxbow lakes are wonderfully developed on large scales for several hundred miles over the flood plain of the lower Mississippi River. The oxbow lakes are there called *bayous.* Oxbow

lakes gradually fill with silt and vegetable matter first to form marshes and finally meadow land.

Tributary Valleys and River Systems. In most cases by far a valley has other valleys tributary to it, and these in turn branch repeatedly into smaller and smaller tributaries. Tributary valleys usually begin as gullies on the sides of the main valley of a region either where the rocks are of uniform hardness, but where rain water moving down the slopes tends to concentrate somewhat more along certain paths than others and hence to erode faster there, or where the materials of the slopes are locally weaker and hence more easily eroded. A gully once started by such a process develops into a valley on the sides of which other gullies form until, under ordinary conditions, a whole system of branching valleys covers a region. A *valley system* thus comprises a main valley with all its tributary valleys, whereas a *river system* comprises a main stream and all its tributary streams. The whole area drained by a river system is called a *drainage basin.*

Fig. 178. A very narrow, steep-sided canyon 1100 feet deep cut in hard rock. Royal Gorge, Colorado.

Accordance of Tributary Valleys. In normal valley and river systems, it is almost always true that a tributary enters a larger or main stream at grade, that is, at the same elevation as the main stream. Such streams and valleys are therefore said to be *accordant.* In a river system which is very actively degrading its valleys it might be presumed that tributaries, with their smaller volumes of water, would not be able to cut down the lower ends of their valleys as fast as the main streams into which they flow. The fact is, however, that, as a main stream sinks its channel, the slope or gradient at and near the mouth of the tributary stream is increased enough to enable the latter, through its

augmented velocity, to cut down as fast as the main stream in spite of lesser volume.

In cases where main valleys have had their sides (especially toward the bottom) cut back and steepened by glacial erosion, or where they have been interfered with by certain other processes, tributaries may enter main valleys at *discordant* levels.

Stages of Valley History. When any new land surface of at least moderate altitude is subjected to erosion by streams, the valleys which develop pass through a life-cycle including stages of youth, maturity, and old age. These stages show certain characteristics by which they can be recognized.

A *young valley* is narrower and steep-sided because down-cutting has thus far greatly predominated over processes of valley widening (Fig. 178). Tributaries are few in number, short, and not well developed. Streams on high lands which are new soon carve out deep valleys with V-shaped cross-sections. Although on newly exposed low lands with gentle slopes the young valleys are, of course, shallow, they are, nevertheless, narrow and steep-sided. Stream velocities are swift.

A *mature valley* is wider, less steep-sided, and usually deeper than a young valley (Fig. 198). It generally has numerous, relatively large, well-developed tributaries. Well along in maturity a flat begins to develop in the bottom of the valley because the stream in it is approaching grade, which means a steady diminution of down-cutting power and an increase in its work of lateral erosion.

An *old valley* shows gently sloping sides, moderate to shallow depth, and fewer tributaries than a mature valley. A wide, nearly level floor (flood plain) also characterizes an old valley because down-cutting by its stream has practically ceased, and lateral erosion has developed the broad flats (Fig. 195, bottom) over which the stream flows slowly in a sweeping, meandering course.

Deposition by Streams

Why Streams Deposit. It should not be assumed from the preceding discussions that streams are everywhere constantly engaged in cutting down their channels and so deepening their valleys. Although the great goal of stream work is to wear down the land to base level, it is nevertheless true that running water does, under certain conditions, deposit sediment. The transportation of sediment by streams into the sea or into inland basins is by no means direct and uninterrupted. Some of the

stream load may be temporarily dropped, whereas some or all of it may be permanently deposited.

The principal causes of stream deposition are (a) diminution of velocity and (b) overloading. It is a law of running water that a partial or complete checking of the velocity of a stream loaded with sediment causes deposition of a part or all of its load. Loss of velocity of a stream may be brought about (1) by decrease of slope of the stream bed,

FIG. 179. The numerous subangular boulders strewn over the desert floor were carried out of the mountains by swift streams, under flood conditions, and deposited because of check in velocity and loss of volume due to water sinking into the ground. Near Desert Center, California.

especially in the lower parts of a large valley; (2) by a decrease in volume, which always means reduced velocity, as when a stream flows through an arid region where loss by rapid evaporation and sinking into the ground is not counterbalanced by new supplies from springs and tributaries; (3) by a change in the shape of the channel, as when a stream enters a wide, crooked channel just after emerging from a relatively straight, narrow channel; (4) by encountering any obstacle such as a boulder or stranded log in a relatively sluggish stream when a sand-bar or even an island may begin to develop; (5) by entering a body of standing water when the current is completely checked, and the whole burden of sediment is dropped. Deposition also takes place where a

swift tributary carries more sediment into a slow-moving larger stream than the latter can carry, or where wind blows into a stream a heavier load than can be carried.

Alluvial Cones and Fans. When a swift, sediment-laden stream emerges at the base of a steep slope from a gully, gorge, canyon, or even ordinary valley upon a more nearly level lowland, there is a tendency for the load to deposit at and near the base of the slope. This is mainly because the velocity of the swift stream is suddenly checked. Such an accumulation of rock débris is generally in the shape of a partial cone. It is called an *alluvial cone* when it is steep, and an *alluvial fan* when its angle of slope is relatively low. Cones and fans vary in width from a few feet to a good many miles, and in thickness from a few feet to many hundreds of feet. They are grandly displayed in the drier portions of the United States at the bases of mountains, as in Utah, Nevada, and southern California (Figs. 180 and 209). An alluvial fan about 40 miles wide has been built by Kings River where it descends from the Sierra Nevada Range upon the nearly level floor of the Great Valley of California.

The shape of the alluvial cone or fan may be explained as follows: At the mouth of the valley of the swift, sediment-laden stream the check in velocity causes deposition of some of the load in the channel, thus choking it, and causing some of the water to spread around the obstruction. The minor streams in turn become choked. Locally some water breaks over the sides of the channels and so develops new channels. By many repetitions of these processes, branching channels, known as *distributaries,* are formed (Fig. 180). Much of the deposition takes place at the mouth of the valley or canyon and directly in front of it, but considerable portions of the sediment spread out on either side, and so the cone or fan-shaped form is developed. A factor which greatly aids the accumulation of the deposit, especially after a cone or fan has been well started, is the decrease in volume of water, and hence in carrying power, due to the sinking of so much of the water into the porous fan material.

Piedmont alluvial plains are made by the coalescence of two or more alluvial fans which are built up by neighboring streams as excellently shown in Death Valley, California (Fig. 207). Another good example is found along the foot of the mountains between Pasadena and Redlands in California where there are many miles of orange groves. In Colorado, Wyoming, and Montana, the rivers which flow eastward out of the Rocky Mountains have developed coalescing alluvial fans into a

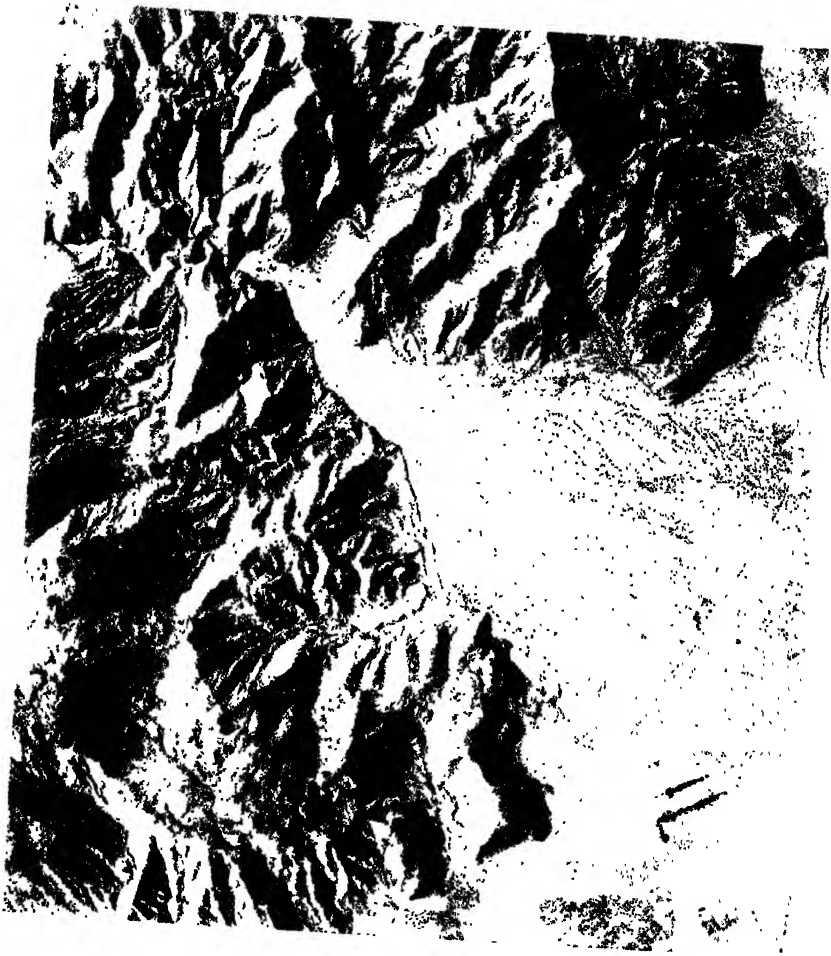


FIG. 180. An aerial view of an alluvial fan in a mountainous desert region. Note the numerous distributaries which are dry channels most of the time. Greatest width of the fan is about two miles. Between nine and eleven miles southwest of Palm Springs, California. (Fairchild Aerial Surveys, Inc.)

practically continuous sheet several hundred miles long, many miles wide, and not uncommonly hundreds of feet thick, from the base of the very steep Front Range eastward upon the Great Plains. The character and structure of the materials, as well as the fossil remains of land plants and animals in them, prove that the deposits were formed on land through the agency of water.

Alluvial cones and fans are most conspicuously developed in mountainous regions with arid or semi-arid climate not only because streams in the mountains are in general much better supplied with water than those upon the lower levels, but also because, as is characteristic of the drier regions, the rain storms may not uncommonly be of the nature of downpours or cloudbursts. Such conditions are very favorable for the development of alluvial cones and fans (Fig. 180). It should not be presumed, however, that alluvial cones and fans are rare in humid regions. In such regions small cones or fans may be seen at the mouths of gullies or small valleys where the soil is very sandy or gravelly, and where much of the sediment-laden water emerging from the gully or valley upon the lowland soaks into the porous material. The courses of some rivers have been notably obstructed as a result of the building of alluvial fans into them by tributary streams. Such a river is forced to flow around the outer border of the fan and over to the opposite side of the valley. A good case in point is Lake Peoria in the Illinois River. In Switzerland the river which emerges from the Lauterbrunnen Valley has built an alluvial fan into and across a long, narrow lake, dividing it into two parts, Lake Thun and Lake Brienz.

FIG. 181. A combination talus slope and alluvial cone. Near Lake Louise, Alberta, Canada.

Stream-bed Deposits. We have already shown how relatively swift tributaries sometimes carry so much material or such coarse rock fragments into the bed or channel of a main stream that a partial dam is built across the latter. This happens because the current of the main stream is not strong enough to keep the material removed. Where

some of the short, very swift tributaries of the Colorado River in the Grand Canyon of Arizona feed such large amounts of coarse rock fragments into the river, the latter, in spite of its swiftness, is not able to remove the fragments fast enough to prevent local near-ponding of its water.

The current of an ordinary stream is so irregular that while, at a given time, much sediment is being moved downstream, some may be deposited in the back water of eddies or in portions of the channel where the current is less rapid.

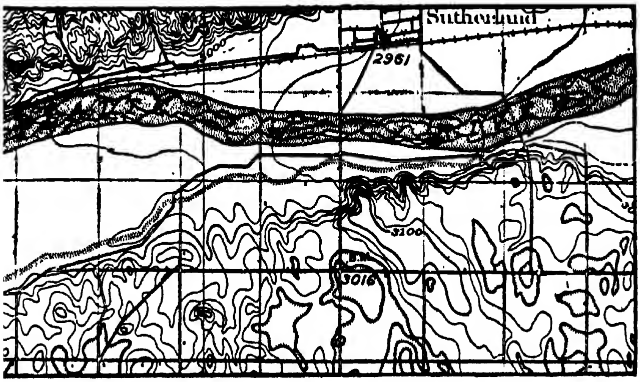


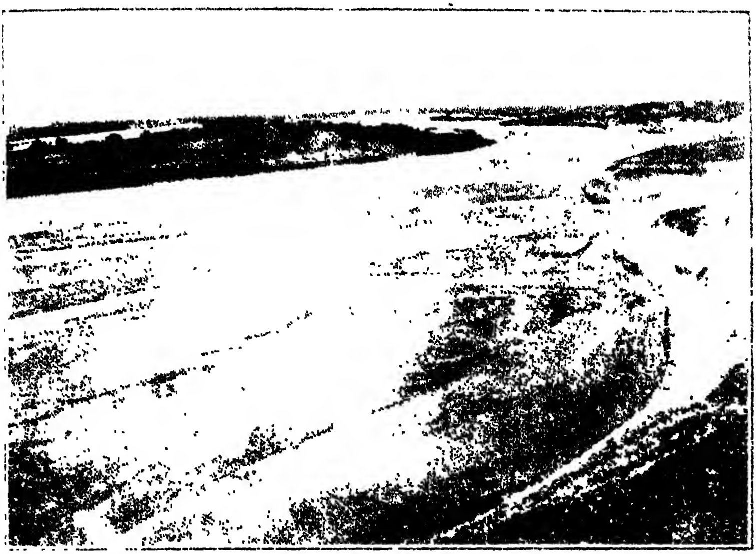
Fig. 182. **A braided stream. Each square represents a square mile. South Platte River, Sutherland, Nebraska.** (After U. S. Geological Survey.)

A stream which is carrying a load of sediment during a flooded condition must, as the flood declines, deposit part of its load in the channel because of loss of volume and velocity. Stream-bed deposits formed in the various ways just mentioned are, however, usually only temporary and of very local extent.

Stream-bed deposits assume much greater importance in the cases of streams which are frequently overloaded and are thus forced to deposit much of their sediment. Such streams build up (aggrade) their beds. An excellent illustration of this principle is the Platte River. Its two main tributaries (North and South Platte) are very swift and, therefore, able to carry unusually heavy burdens of sediment in times of flood. These two tributaries, as well as the main stream formed by their confluence, lose much of their velocity and considerable volume through evaporation when they emerge from the mountains and flow out upon

the drier and much more gently sloping Great Plains of Nebraska. In spite of these conditions, the stream load is rather effectively carried during high water, but, when the flood subsides, considerable deposition takes place on the river bed because the stream then becomes more or less overloaded after emerging from the mountains.

Much of the time the Platte River of western Nebraska is a ***braided stream,*** that is, one which does not flow in a single definite channel but



;. 183. Braided channel of a sediment-laden stream. North Platte River on Nebraska-Wyoming state lien. (After Darton, U. S. Geological Survey.)

rather in a network of ever-changing, branching and reuniting channels (Fig. 182). The local portions of the stream flowing in such channels are called *distributaries.* They are easily explained as follows: When sediment is deposited on the bed of a channel the latter becomes too small to hold all the water, part of which then breaks over the side and flows in a new course. When the new channel becomes sufficiently clogged it in turn develops branches. By many repetitions of such a process and the frequent reuniting of channels, the network of courses of a braided stream is produced. The braided course does not exist as such during high water because then the whole flat, which during lower water contains the network of channels, is covered by the stream.

Gravel or sand bars form in some streams which do not become braided. These are most likely to develop at times of low water. A given bar may be partly or wholly cut away by high water (with increased velocity) or it may last for some time as a low-water island.

Meander Deposits. When a stream reaches a graded or approximately graded condition and develops meanders, it does so by a twofold process of cut-and-fill. While the current which is directed against the bank of the outer portion of a meander is there performing the work of lateral planation the current is relatively slack on the side of the channel directly opposite, and so deposition takes place there up to flood level. On the side where cutting takes place, the bank is steep and the water deepest, while on the opposite (filling) side the bank slopes gently and the water is shallowest. If it were not for this twofold process of filling on one side of the channel and cutting into the opposite bank, the meander could not long continue to develop because cutting alone would widen the channel to such an extent that the greatly weakened current would lose its power of lateral erosion (Fig. 185).

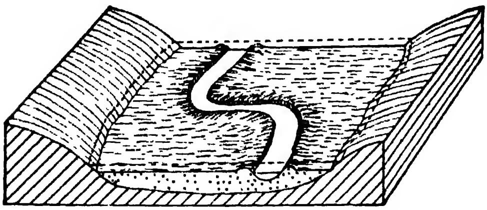


FIG. 184. Diagram illustrating river flood plain, deposits, and natural levees. Dotted line shows high-water level.

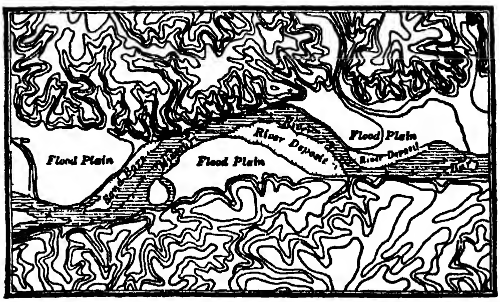


FIG. 185. Sketch map showing an early stage of the meandering of a stream. (From Tarr's *New Physical Geography* by permission of the Macmillan Company.)

Flood-plain Deposits. In most cases by far flats on valley bottoms are developed by the lateral erosion of streams, particularly when they are graded or nearly so. This process has already been explained. In some any land area with its valley subsides so much that enough deposition of cases valley-bottom flats are formed by aggradation, as is the case when

sediment must take place in the valley to build up its floor to a graded condition. However they are formed, valley flats subject to overflow during high water are called *flood plains,* defined on page 203.

When a typical flood plain is covered by high water, the current following the main (low water) channel has its velocity greatly augmented so that not only is its power to transport increased, but also it then actually erodes (cuts down) its channel. In the meantime the sediment-laden water over the flood plain has a velocity much less than that of the main current so that some deposition takes place there (Fig. 184). It should not be inferred from this that flood plains always build up very much through deposition because, if a stream is practically graded and the land is not subsiding, the shifting of meanders back and forth all over the flood plain from one side to the other keeps the average level of the plain from changing very much.

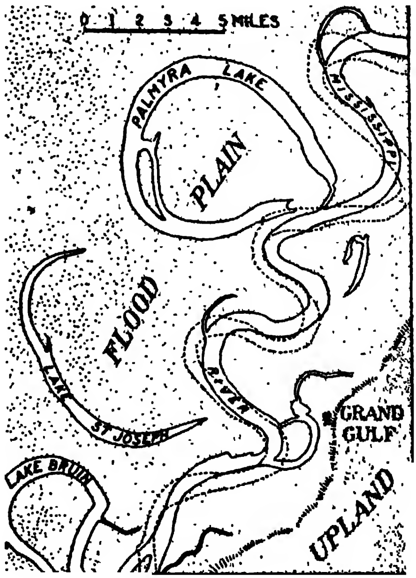


Fig. 186. Meanders and oxbow lakes of part of Mississippi River flood plain in 1883 (heavy lines) and 1896 (dotted lines). (By William Davis, based upon Government Surveys.)

When muddy water covers the flood plain of a river, the conditions for deposition are most favorable along the edges of the main channel because there the sediment-laden current of the swift-moving channel water is suddenly greatly checked by friction against the slower moving waters of the flood plain. Because of this sudden diminution of carrying power along the edges of the main channel, more and coarser materials deposit there than over the general flood plain surface. Low ridges of such origin are called *natural levees* (Fig. 184). Such levees can build up only to higher flood levels, and so they cannot, in the course of time, keep high waters from overflowing the flood plain. They are often built up artificially to prevent streams from overflowing their flood plains.

Some Great Floods. The portion of the lower valley of the great Mississippi River which is subject to floods (i.e., the flood plain) covers an area of about 30,000 square miles. It reaches from above the mouth of the Ohio River to the Gulf of Mexico—an air line distance of 600 miles or about twice that far if measured along the meandering course of the stream. On account of the richness of the alluvial soil many people live on this vast flood plain in spite of the fact that wide portions of it are subject to more or less disastrous floods. Thus in 1903 a portion of the flood plain located in the state of Mississippi became inundated, driving 65,000 people from their homes, flooding half the city of Greenville (population 8000), suspending traffic for 20 days, and causing great damage to property. The floods of 1881, 1882, 1884, and 1897 on portions of the Mississippi River flood plain were particularly disastrous, that of 1897 covering thousands of square miles and causing great inconvenience and a property loss of many millions of dollars to over 50,000 people. Still worse was the great flood of 1912 in the lower Mississippi Valley when the cities of Memphis, Vicksburg, and New Orleans all suffered severely. Fully 30,000 people were rendered homeless north of Louisiana, and over 100,000 in that state lost much property. It has been estimated that this single flood caused a direct loss of fully $75,000,000.

The famous Dayton, Ohio, flood of 1913 was caused by five days of heavy rainfall over the Miami and Scioto River Basins, on ground that was already soaked. More than 200 towns were more or less inundated, and over 400 people lost their lives. Dayton, which is built mainly upon the flood plain of the Miami, suffered most. The flood passed over levees more than 20 feet high and covered much of the city with water 10 feet deep. The property damage was about $30,000,000. At the same time great damage was done in the city of Columbus situated on the Scioto River. Although many hundreds of miles of artificial levees have been constructed to control more or less effectively much of the flood water of the Mississippi, nevertheless the river not infrequently breaks through portions of the artificial embankments and floods local portions of the great flood plain.

The famous Passaic River flood of 1902 in northern New Jersey caused damage to the extent of millions of dollars, especially in Passaic and Paterson.

The disastrous Johnstown, Pennsylvania, flood of 1889 was due to the giving way of a dam as a result of heavy rains, and over 2000 people were drowned. The examples given are only a few of the more destructive of the many river floods which have affected various parts of the United States during the last half-century.

Probably the most awful river floods of known human history have been those of the Hwang-ho ("China's sorrow") of China. The vast flood plain, which is really a delta (see beyond), is 400 miles long and from 100 to 300

miles wide. Its fertile soil has been densely populated for many hundreds of years. A few of the more recent greatest floods took place in 1820, 1858, 1887, and 1892. In each case many villages were wiped away and great numbers of people were drowned. One of the most terrible floods was that of 1887 when more than 1,000,000 people lost their lives either through drowning or starvation, and hundreds of villages were destroyed. In 1892 the mighty river shifted its course during a flood on the delta flood plain to such an extent that its mouth was about 300 miles farther north after the flood subsided. Just before the flood the river emptied into the Yellow Sea (Fig. 187). Ever since the flood it has emptied into the Gulf of Pechili. A number of shifts back and forth from the Sea to the Gulf have occurred within the last few thousand years.

Deltas

Cause of Deltas. Much sediment carried by a stream finally reaches its mouth. If the stream flows into a lake or the ocean, the velocity of the current is largely or wholly checked, and thus much or all of the sediment must be deposited. The destination of most streams is the sea, and, where tides or shore currents in the sea are relatively weak, the discharged sediments accumulate mainly at and near the mouths of the streams in the form of flat, partly submerged, fan-shaped deposits called *deltas.* The name has been given because of the crudely triangular shape similar to the fourth letter of the Greek alphabet. If there are strong tides or shore currents in the body of water which the stream enters, or if the amount of sediment discharged by the stream is relatively small, the tendency is for the sediment to be swept so far away from the mouth of the stream that either no delta will form or only a small or imperfect one will develop. Another reason for the absence of deltas from the mouths of many existing rivers (even some large ones) is the sinking of the land, causing notable submergence of the mouths of the rivers so recently that there has not been time enough for the discharged sediments to build up real delta deposits around the newly located mouths.

Examples of Deltas. Some examples illustrative of the principles just explained will now be given. Very large and typical deltas have been, and are being, formed where big rivers empty into certain lakes or nearly enclosed arms of the sea. Thus the great Nile River has built into the Mediterranean Sea a very typical delta covering about 10,000 square miles (Fig. 188). The Mississippi River has extended its delta

of 12,000 square miles some 200 miles into the Gulf of Mexico (Fig. 189). Extensive deltas have been built by the Danube River into the Black Sea, and by the Volga River into the Caspian Sea. In the face of considerable tidal action, the Hwang-ho River of China has built into the Yellow Sea a vast delta of fully 100,000 square miles (Fig. 187). The combined Ganges and Brahmaputra Rivers of India have, in spite of very considerable tides, formed a delta covering fully 50,000 square miles. The two last-named deltas have formed in spite of rather adverse tidal conditions because of the unusually great quantities of sediment discharged by the rivers. Deltas of even moderate size are absent from the Atlantic Coast of North America, not only because of strong tidal action, but also because of notable subsidence of the land very recently (geologically considered) with resultant submergence ("drowning") of the lower portions of the rivers. This is strikingly illustrated by the St. Lawrence and the Hudson Rivers.

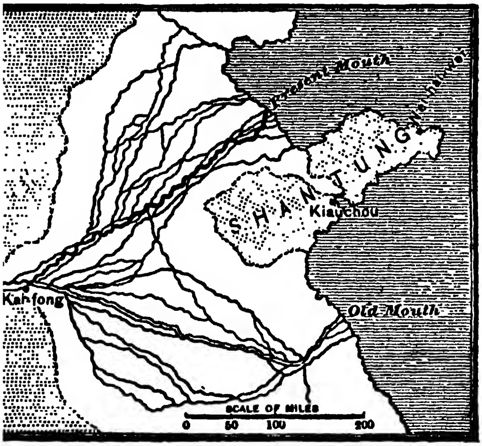


FIG. 187. Delta of the Hwang-ho River, China. (From Tarr's *New Physical Geography* by permission of the Macmillan Company.)

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The Delta Surface. What are some of the characteristic features of the common type of delta? Its surface is a wide, nearly flat, usually fan-shaped plain mostly a little above and partly a little below the level of the body of water into which it grows. Thus about two-thirds of the surface of the Mississippi delta is above water under ordinary conditions, but most of it is inundated by high water during a flood. The great bulk of delta material is, however, always under water, and thus it differs from an alluvial cone or fan whose material is wholly or largely on land. Another almost universal feature of a delta surface is the presence of *distributaries,* that is, branches into which the stream splits in increasing number, beginning at the head (upper end) of the

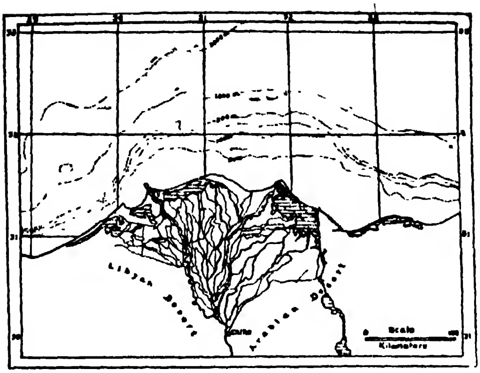

FIG. 188. Delta of the Nile. Depth of water in meters. (After J. Barrell.)

FIG. 189. Map of part of the Mississippi delta about the year 1885. Width of map area, 106 miles. Depth of water in fathoms. (After U. S. Coast and Geodetic Survey.)

delta. These distributaries wander over the delta plain in an ever-widening network, and so a delta-forming river always has several or many mouths (Figs. 188 and 189).

Delta Structure. The delta shows a characteristic structure because of the special conditions under which deposition of the sediment takes place. Thus the steep front (Fig. 190), so characteristic of a delta, results from rapid deposition of the coarser sediment layer upon layer where the onrushing sediment-laden stream (or each mouth of the stream) meets the relatively deeper standing water into which the stream flows. These steeply inclined layers are called *fore-set beds* (Fig. 190). They make up the greater bulk of the delta pile. The

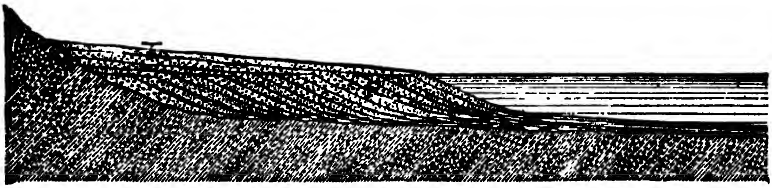

FIG. 190. Ideal structure section of a delta. T = top-set beds; B = bottom-set besd; F = fore-set beds. (Modified after G. K. Gilbert.)

finer sediments spread out in layers over the floor of the lake or sea to a greater or less distance out from the base of the steep front of the delta. These layers are called the *bottom-set beds.* The earlier formed bottom-set beds of course become buried under the fore-set beds. The *top-set beds* are deposited by the stream on top of the fore-set beds as the latter advance into sea or lake and shoal the water. They build up, for the most part, to a little above the level of sea or lake in layers which slope very gently seaward or lakeward.

Rate of Growth of Deltas. Some rather accurate data regarding the rate of growth of various deltas are known. A few examples will be mentioned. One mouth of the Mississippi River is growing into the Gulf of Mexico at the phenomenal rate of one mile in 16 years (Fig. 191). The River Po has extended its delta 14 miles into the Adriatic Sea in 1800 years as proved by the fact that Adria, a seaport at the mouth of the river, 1800 years ago, is now 14 miles inland. The Rhone River has been building its delta into the Mediterranean Sea at the rate of one mile in 100 years for many centuries. The ancient seaport of Rome is now three miles inland because of delta extension by the Tiber

River. But by no means do all deltas build out so fast. Thus the great delta of the Nile has grown seaward but little in 2000 years because a current sweeping along the delta front is strong enough to keep the sediment removed about as fast as it is supplied by the river. The Amazon River has not been able to build a delta deposit even up to sea level because of the very strong tides and sea waves, though it has constructed an extensive submarine delta covered by water less than 60 feet deep.

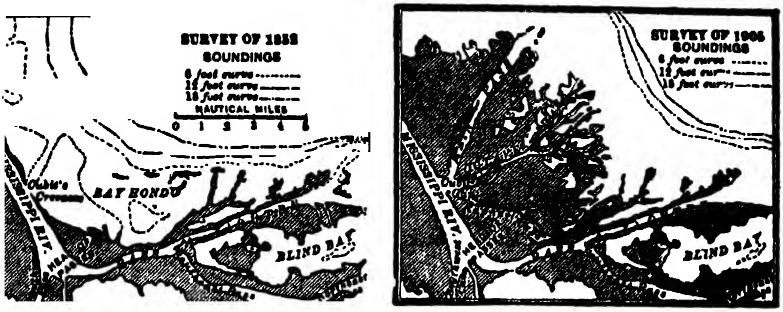

FIG. 191. Maps showing growth of part of the Mississippi delta. (After Putnam, from Tarr's *New Physical Geography* by permission of the Macmillan Company.)

Subsidence of Deltas. Many of the great deltas of the world have been slowly subsiding while deposition of sediment has been going on. Where the rate of deposition has been somewhat faster than the rate of sinking, typical deltas have developed, but where subsidence has been predominant, even with other conditions favorable, deltas have not been built up above sea level. Thus the Ganges, Nile, and Mississippi Rivers have built up very extensive deltas in spite of subsidence of fully hundreds of feet. This has been proved by borings into the deposits. By this method "layers of peat, old soils, and forest grounds with the stumps of trees are discovered hundreds of feet below sea level. In the Nile delta some eight layers of coarse gravel were found interbedded with river silts, and in the Ganges delta at Calcultta a boring nearly 500 feet in depth stopped in such a layer" (W. H. Norton). These are of course top-set beds which have, along with the underlying fore-set and bottom-set beds, subsided. Deltas of any consequence are absent from the middle Atlantic coast of North America because of very recent sinking of the land, causing the development of *estuaries* by flooding of the lower ends of the river valleys with tide water.

History of Stream Courses

Consequent Streams. On any new land surface, the first streams will have their courses determined by the original slope and natural irregularities of the surface. Such stream courses are, therefore, consequent upon the original relief features. They may, of course, not only lengthen by headward erosion and deepen and widen their valleys, but also they may have tributaries developed as a direct consequence of the initial topography. All such streams whose courses are the direct consequence of the initial topography are called *consequent streams.*

New land surfaces may develop in various ways, some of which will now be very briefly explained. A portion of the sea floor may be raised into land with a relatively smooth surface sloping seaward. Examples are the outer portions of the Atlantic and Gulf Coastal Plains of the United States which are of geologically recent origin. In this region the southern part of Florida is so recent that its consequent drainage is exceedingly young. If a portion of the crust of the earth is newly upraised into a ridge or range by an earth-crust disturbance, consequent streams develop courses down each side of the ridge, as is the case with the Sierra Nevada Range. Where the newly uplifted mass is dome-shaped, or where a new volcanic cone develops, the consequent streams radiate downward in all directions from the summit. Newly built-up lava plateaus, like those in Yellowstone Park, eastern Washington, or southern Idaho, will have consequent streams developed upon their surfaces. Where new land surfaces result from widespread deposition of materials by glaciers, especially vast glaciers like those of the Ice Age (Fig. 249), over older land surfaces, consequent streams develop on the new surfaces. Large portions of Iowa and Illinois are good illustrations.

Subsequent Streams. During the history of a drainage system, it happens almost invariably that many stream courses originate independently of the original (initial) topography and are determined and regulated by erosion proceeding differently upon the bedrock formations according to differences in hardness, structure, and resistance to erosion of the formations. Such streams are said to be *adjusted* because they carve out their valleys along lines or belts of the weaker or more yielding rocks. During the progress of erosion of a region, the divide (division of drainage) between two streams, with courses in rocks of like character, often shifts position notably by headward erosion of the stream with the greater power to erode toward the one with the lesser power,

and the upper course of the former stream is not consequent. Also during the history of a drainage system, streams (or portions of them) are not uncommonly captured by (drained off into) other streams, thus bringing about changes in stream courses not consequent upon the original topography of the region.

All streams which develop independently of, and subsequent to, the original relief of a land area, whether by adjustment to rock character or structure, shifting of divides, stream capture, or any other process,

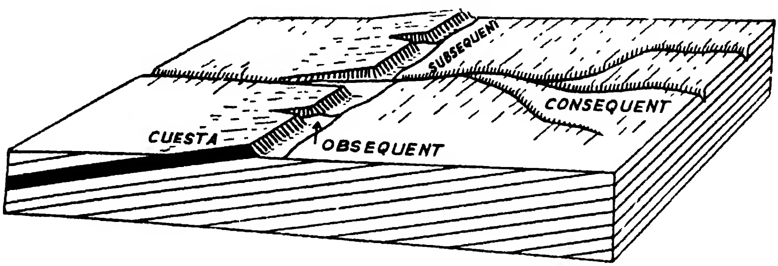

FIG. 192. Block diagram to illustrate the meaning and relations of consequent, subsequent, and obsequent streams, in a youthful stage of the normal cycle of erosion, in a region of tilted strata. The more resistant rock layer (in black) stands out in the form of a cuesta, against the erosion.

are called *subsequent streams* in distinction from consequent streams. Subsequent streams are very commonly tributaries of consequent streams, but even a consequent stream course may, during the progress of erosion of a region, undergo sufficient change to become subsequent. "The original consequent course may, for example, be very irregular and roundabout, and during its development such a course will tend to be straightened. Or the original course may be straight, and, with subsequent development, meandering will be set up. . . . Usually subsequent tributaries develop at (or nearly at) right angles to a consequent master stream. This angular pattern of stream courses is known as *trellis drainage*. . . . Streams in which no adjustment to rock structure takes place, either (a) because of widespread flat-lying sediments, or (b) because the stream develops in a large area of a massive formation such as granite, never have subsequent tributaries. This is because the adjustment is complete from the beginning. This *insequent* stream pattern is often tree-like, for which reason the drainage is said to be *dendritic*" (Tarr and Martin).

Obsequent Streams. A subsequent stream which develops as a tributary at about right angles to a consequent stream may in turn have tributaries. "Those which flow in the opposite direction from the original consequent drainage are called obsequent streams" (Tarr and Martin).

Conditions particularly favorable for the development of obsequent streams are to be found in regions of tilted strata where escarpments, such as cuestas, form as the land surface is worn down by erosion. In such cases obsequent tributaries will develop on the escarpments as shown in Fig. 192.

Normal Cycle of Erosion of Regions

Introduction. In the discussion of the stages of valley history it was pointed out that an individual valley, in due course of time, may pass through a cycle of youth, maturity, and old age. Just as surely as that is true for the individual valley and its stream, so does an extensive topographic region, comprised of many valleys and stream systems and of intervening divides and uplands, undergo a series of topographic and drainage transformations with the passage of time. The total series of changes in the configuration of a land surface is referred to as *a cycle of physiographic development* or, perhaps better, *the cycle of erosion*. Using a biological analogy, the most important stages of such a cycle have been called *infancy, youth, maturity,* and *old age*. A cycle of erosion may be defined as "the period of time during which an uplifted (or any new) land mass undergoes its transformations by the processes of land sculpture (erosion), ending in a low featureless plain." These transformations may be relatively simple and easily understood or they may be very complicated because of wide variations of contributing causes such as climate (especially rainfall), altitude of the land, character and structural relations of the rocks, diastrophic interferences, and others.

By a *normal cycle of erosion* we mean the time required for the reduction to or near base level by erosion (mainly stream action) of a new land area of at least moderate altitude with a humid climate and with no interfering change of level of the land by earth-crust movements. The principles involved may be best set forth by tracing through the stages in the topographic development of a land mass under the conditions just described and with an initial, smooth, sloping surface reaching to hundreds or even thousands of feet above sea level. Beyond

in this chapter, some of the more important variations and interferences with the so-called normal cycle are discussed.

Infancy Stage. A typical newly formed land surface, like the kind just pictured, has a drainage system developed upon it. In the earliest stage (*infancy*) of its cycle of erosion, only a few streams form, and these tend to seek out the original depressions and to flow down the

Fig. 193. An aerial view of gully development illustrating a region in topographic youth. Kettleman Hills, California. (© Spence Air Photos.)

initial slope of the land. These are, of course, consequent streams. From the very start some of these streams will be longer, larger in volume, and more energetic as erosive agents than others. A characteristic of all is the small number of tributaries. Not uncommonly, some original basin-like irregularities or depressions will be filled with water to form ponds or lakes. During infancy, stream erosion accomplishes very little, but the process of sheet erosion is then most effective.

Youthful Stage. The region relatively soon passes into the next stage called *youth* (Fig. 195). During this stage the streams carve out narrow, very steep-sided valleys usually with V-shaped cross sections

(Fig. 194). All the streams are very actively engaged in deepening their valleys by erosion or, in other words, none of them has reached a graded condition. Flood plains and meandering streams are therefore lacking. During this youthful stage, there are few if any sharp divides (divisions of drainage), and the streams are still relatively few in number. The relief of the region is, for the most part, not rough. General

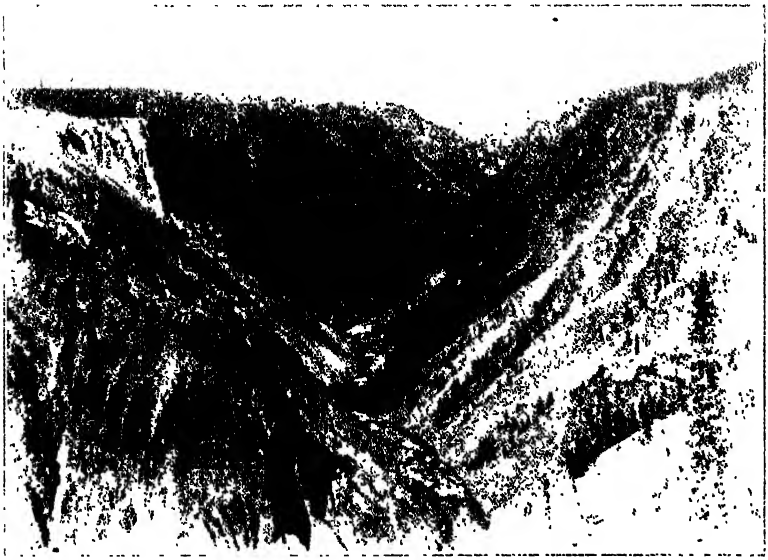

FIG. 194. A youthful V-shaped valley cut in a high plateau of lava. Grand Canyon of the Yellowstone, Yellowstone Park. (After W. T. Lee, U. S. Geological Survey.)

erosion of the region is not far advanced because erosion is largely confined to the relatively few stream channels. Gorges (or canyons), waterfalls, or lakes (or ponds) are not uncommonly present because they are geologically short-lived features which are indicators either of a youthful topographic stage of a region or of some geologic process or disturbance which has recently affected a topographically older region, as pointed out beyond. Some examples of regions in topographic youth are much of the Atlantic Coastal Plain of the United States which has recently emerged from the sea; the Colorado Plateau of Arizona, with its Grand Canyon, which has been upraised, in late geologic time, to its present altitude; and the general vicinity of Fargo, North Dakota,

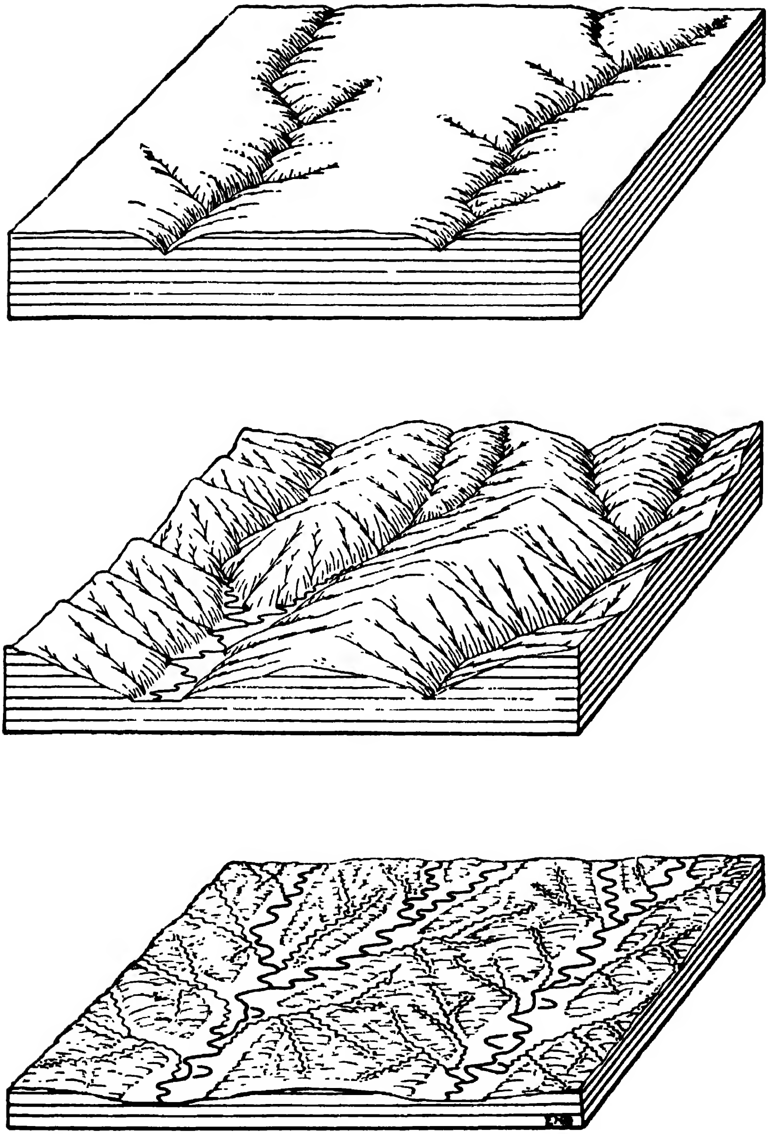

FIG. 195. Block diagrams to illustrate the normal cycle of erosion in a moderately elevated region of essentially uniform, stratified rocks devoid of important structures. Top, region in youth; middle, in maturity; bottom, in old age. See text for characteristics of stages.

which is part of a large lake bed from which the water has been so recently drained that it is in early youth.

Mature Stage. Erosion continues until the features so characteristic of youthful topography gradually give way to those distinctive of *maturity* (Fig. 195). A region in typical maturity has the maximum number (intricate pattern) of streams most of which flow in valleys which are wider and less steep-sided than those of youth, that is, their cross sections are broader V-shaped. The maximum roughness of relief has

Fig. 196. A mountainous region in early maturity where resistant granite rocks prevail. Western San Gabriel Mountains, California.

developed (Fig. 195). Divisions of drainage (divides) are well defined and sharp. General erosion is, in fact, then most effective because practically the whole region has been cut to steep slopes. Waterfalls, gorges (or canyons), or lakes (or ponds) rarely if ever exist because enough time has been given for the streams to obliterate such temporary features. Between middle and late maturity one (or more) of the larger streams of the region has cut down near enough to a graded condition (at least in its lower course) to begin meandering with resultant development of a flood plain. During maturity a river system does its maximum work in regard to amount of down-cutting, quantity of water carried off and load of sediment transported. A very fine illustration of a region in typical maturity is that around Charleston, West Virginia. A wide region around Lancaster in southwestern Wisconsin, including a part of the Mississippi River, is also a good example of

maturity. Except for the very recent addition to it here and there of some minor features of youth, as inheritances of the Ice Age, such as gorges, waterfalls, and ponds, the region of central-western Massachusetts, with its Connecticut Valley, is a fine example of a region in late maturity (Fig. 198).

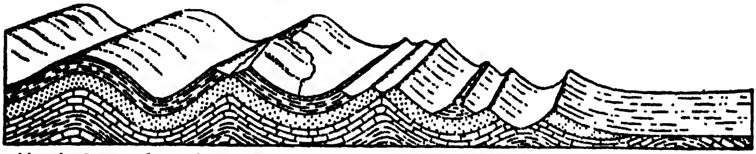

FIG. 197. Diagram to illustrate successive stages in the normal cycle of erosion in a region of folded rocks. (After G. H. Ashley.)

Old-age Stage. As the erosion of the region progresses, the *old-age* stage is reached when the relief has been greatly subdued to the condition of an undulating plain, or so-called "rolling country." Divides are then not at all sharp, being low, rounded hills. Only a moderate number of streams remain, and these flow through wide, shallow valleys. Most of the streams (especially the larger ones) are graded or nearly so, and their sweeping meanders and cut-off meanders (oxbows) on wide flood plains are common and characteristic. General erosion and the amount of work accomplished by the streams are much less than in maturity. Gorges and waterfalls are absent, but oxbow lakes, which are easily distinguished as such, are present. A region well along in the old-age stage of its erosional history has a highly subdued topography of very low relief, which, in the final stage of a perfect

FIG. 198. Maturity in an area of varied rocks and structures. Connecticut Valley region of Massachusetts. (Howell's relief model.)

cycle of erosion, would be a featureless plain at base level. It is doubtful if any wide area has ever been reduced to such a base-leveled condition, although such a condition has often been rather closely approached. Very typical examples of old-age topography are south-central Kansas in the general vicinity of Caldwell, and the region around Butler, Missouri.

Time Involved. The terms infancy, youth, maturity, and old age, as above employed, do not denote anything like definite periods of years, but rather they represent stages, each with certain characteristics, of the cycle of erosion of any given region. Since conditions of altitude, slope, rock character, and rainfall of different regions vary so widely, it is clear that either topographic maturity or old age may, as measured in years, be reached in one region or even a portion of a single region long before it is in another. The terms under consideration "have reference not so much to the length of their history in years as to the amount of work which streams have accomplished in comparison with what they have before them."

Peneplains and Monadnocks. *Definitions.* Any region which has been worn down by erosive agencies to a condition of very low relief at, or nearly at, base level is called a *peneplain* (or peneplane), meaning

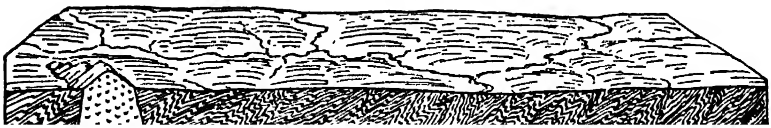

FIG. 199. Block diagram showing a peneplain surmounted by a monadnock of more resistant rock. (After W. M. Davis.)

an "almost-plain." A perfect peneplain would be a plain wholly at base-level-of-erosion, but because probably no large land area has ever been completely base-leveled, it is customary to call a region of faint relief, well along in the old-age stage of its erosion, a peneplain. Perfect base-leveling of a region must rarely if ever take place because, as old age is approached, the rate of erosion becomes slower and slower as the gradients of the streams become less and less, so that the time necessary for perfect planation would be almost infinite. Almost invariably diastrophism, igneous activity, glaciation, or some other process notably affects the region long before anything like a perfect peneplain or base-leveled condition is reached.

It often happens that, during the development of old-age (or peneplain) topography, more or less local portions of the region are not cut down to the general peneplain level, either because they consist of more resistant rocks or because they lie in the midst of relatively wide spaces between larger streams and so are more favorably situated against erosion. Such a residual mass rising well above the general peneplain level is called a *monadnock,* so named after Mount Monadnock of New Hampshire which rises conspicuously above the now upraised and partly eroded peneplain of southern New England.

Existing peneplains. Peneplains, or even reasonably close approximations to them, are not very common over wide areas of North America, as may be inferred from the above discussions. One reason for this is the fact that so many portions of the continent, including many well-worn-down (practically peneplaned) areas, have been more or less uplifted in recent geologic time (Cenozoic era) and subjected to renewed erosion. Much of the large area comprising central-western Missouri, southeastern Kansas, and northeastern Oklahoma has been reduced by erosion to a condition of old-age topography approaching a peneplain, though still lying hundreds of feet above sea level.

Recently upraised peneplains. The vast eastern Canadian region, extending from near the international boundary northward to either side of Hudson Bay, consists of a complex mixture of very old rocks and structures which, after long ages of geologic time, was mostly reduced to a common level of very low altitude. This so-called "Laurentian Peneplain" has since (in the Cenozoic era) been rather unevenly upraised in amounts varying from a few hundred feet to about 2000 feet. In the interior of Labrador, for example, the old, eroded, upraised, peneplain surface is so smooth that variations of level are only a few hundred feet within an area of 200,000 square miles.

It has long been known that most of the eastern United States, where the higher lands such as southern New England, New York, and the northern and central Appalachians are situated, was subjected to such widespread and deep erosion that, by middle Cenozoic time, it was largely reduced to the condition of a remarkably smooth plain (peneplain) near sea level with slow-moving (graded) streams meandering over its surface. In reference to this great peneplaned area Berkey has said: "The continent stood much lower than now. Portions that are now mountain tops and the crests of ridges were then the constituent parts of the rock floor of the peneplain not much above sea level. . . . The ridges and valleys, the hills, mountains, and gorges of

the present were not in existence, except potentially in the hidden differences of hardness or rock structure." In southern New England, and in the eastern Adirondack Mountains of New York, some monadnocks stood out above the peneplain level. Mounts Greylock and Monadnock of southern New England mark the sites of such remnants of erosion (monadnocks). The differential uplift of this vast peneplain to altitudes up to a few thousand feet took place in the latter part of the present (Cenozoic) era. It would carry us too far afield to go into the various proofs for the existence of this once great peneplain. One important line of evidence may, however, be mentioned, namely, the so-called "even sky-line." Since rocks of many kinds and ages were, with very few exceptions (the monadnocks), all truncated or planed off to a general level as indicated by the "even sky-line," it is impossible to account for the present topography except as an upraised and partly eroded peneplain (Fig. 204).

Interrupted Cycles of Erosion. *Rejuvenation.* The normal cycle of erosion, which, as we have shown, ends with a peneplain condition, may be interrupted at any stage by other processes. Such processes are so varied, and their effects are often so complicated, that we shall attempt only to explain briefly some of the more important ones and their general effects.

Fig. 200. A rejuvenated region showing entrenched meanders. Yakima Canyon, Washington. (Hobbs, after G. O. Smith.)

The most common and significant cause of interruption of the normal cycle of erosion is change of level of the land (diastrophism). Thus a region in any stage of its erosional history prior to almost complete peneplanation, say in maturity or early old age, may be upraised with

resultant notable increase in velocities of the streams. Such a region is said to be *rejuvenated,* and the streams whose activities are quickened are said to be *revived.* The revived streams begin to cut youthful valleys in the bottoms of the wider, older valleys, and thus a new cycle of erosion is inaugurated. The effect is at first most pronounced on the valley floors of the streams which are graded or nearly so, but in time the whole region is distinctly affected by the revival of erosive activity.

FIG. 201. A rejuvenated region showing a youthful valley cut into an older, mature valley. Matanuska Valley, near Glacier Point, Alaska. (Photo by Mendenhall, U. S. Geological Survey.)

If, through processes of erosion, a meandering stream develops on a flat valley-bottom, and then uplift of the land takes place, the revived stream proceeds to cut a youthful valley in the old valley floor without changing its meandering course. Such meanders are known as *entrenched* (or *incised*) *meanders.* Among numerous excellent examples are the San Juan River of southeastern Utah, Yakima Canyon in Washington (Fig. 200), the Susquehanna River of southern New York and northern Pennsylvania, and certain rivers of western Germany, Belgium, and northwestern France.

In a case of uplift of a region which has undergone practically complete peneplanation, or base leveling, the general effect is, in nearly all respects, like a new land surface (with consequent streams) formed in

other ways, and it may be treated as such. Such a case scarcely comes under the category of an interrupted cycle of erosion.

Rejuvenation by uniform uplift. Rejuvenation may be by uniform uplift of an old eroded surface, in which case the altitude is increased, but the attitude or slope of the surface is not changed. An excellent case in point is the area of thousands of square miles of central and southwestern New York where many remnants of an upraised, eroded (peneplaned) surface lie at a remarkably even altitude of about 2000 feet,

FIG. 202. An old river channel cut in bedrock and later filled with waterworn boulders, as seen in a road cut hundreds of feet above the bottom of a present-day valley. Because of rejuvenation by uplift, erosion is again active in cutting away the old river deposit. Five miles west of Reno, Nevada.

thus indicating a practically uniform uplift of the old eroded surface to its present height. This upraised peneplain has been deeply and widely trenched by erosive processes, including the development of the Mohawk Valley.

Rejuvenation by tilting. Tilting (without faulting) may accompany uplift of an old eroded surface. This is well illustrated in the case of southern New England where a fairly well-developed peneplain was upraised with a distinct tilt southward, as indicated by the slope of the "even sky-line" of the numerous remnants of the old eroded surface.

Another example is the southwestern part of the Colorado Plateau which is an old, approximately peneplaned surface upraised thousands

of feet, with a southward downslope, and deeply trenched by the Colorado River.

Rejuvenation by warping. An erosion surface may be more or less warped during uplift. This principle is finely illustrated by the central Appalachian district. Millions of years ago, the strata were deformed by folding. Then the region was cut down by erosion to an almost perfect peneplain which, in relatively recent geologic time, was distinctly upwarped along a north-south axis and then deeply dissected by erosion. Plainly preserved remnants of the upwarped peneplain do not, therefore, rise to a uniform altitude, but they rise steadily higher as the axis of upwarp is approached from both the east and the west sides.

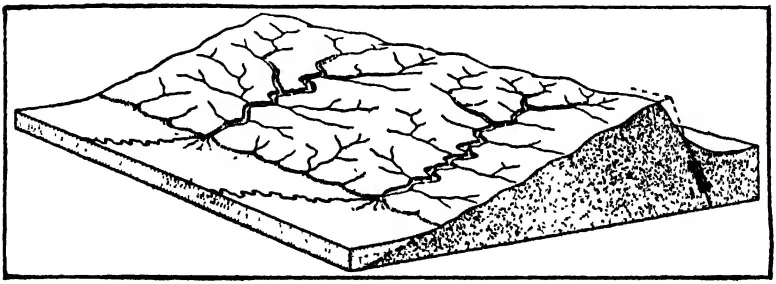

FIG. 203. Diagram representing a portion of the great Sierra Nevada fault-block rejuvenated by faulting in later Cenozoic time. (After Matthes, U. S. Geological Survey.)

Rejuvenation by faulting. Still another case is the interruption of a cycle of erosion by faulting as wonderfully illustrated by the Sierra Nevada Range of California. The site of the range was once in the topographic condition of late maturity or early old age, and stream gravels were spread over the old, eroded surface in many places. Then a profound fault fracture developed along the eastern side of the district, and the great Sierra earth block was upraised many thousands of feet and tilted at the same time towards the west (Fig. 203). The long, wide, western face of the range, therefore, shows a fairly "even sky-line," but by no means a level one for it gradually descends from a summit altitude of 7000 to over 14,000 feet westward, nearly to sea level. Plainly preserved remnants of the above-mentioned stream gravels scattered over the western slope of this vast fault-block range prove that we are here dealing with an old, eroded surface uplifted and tilted by faulting.

Cycle interrupted by subsidence. Subsidence of the land also interferes with a normal cycle of erosion. Its general effect is to hasten old age by diminishing the amount of erosive work the streams have to do. Continued sinking causes deposition of sediment in the valleys (or portions of them), thus building up their floors. "If it can be proven that all the graded streams of a region have their beds at levels far above beds which they previously occupied, it seems most likely that the sur-

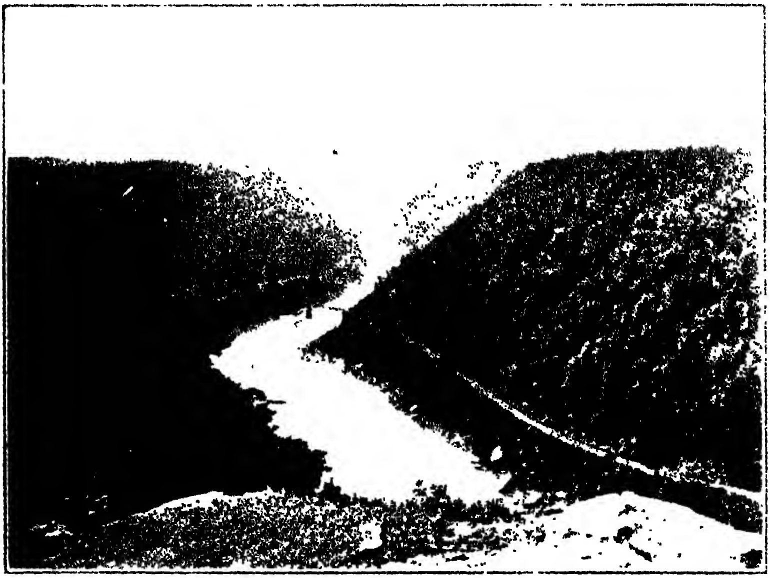

Fig. 204. A valley being cut in the uplifted Appalachian peneplain. Note the even sky-line of the mountains. New River, Virginia. (After Hillers, U. S. Geological Survey.)

face of the region has subsided. . . . If uplift raises a previously graded surface above grade, subsidence lowers valley bottoms below the level of grade. This principle seems to be illustrated in the upper Mississippi Valley region, where the Mississippi River and its main tributaries are at grade 100 feet or more above the bed rock beneath" (Trowbridge). The fills in these valleys consist of loose glacial and water-laid sediments.

If a seaboard region in any stage of erosion, particularly from typical youth to typical old age, subsides enough relative to sea level, tidewater floods at least the lower courses of the valleys and their streams, and they are said to be *drowned* (Fig. 41). Not only are such valleys, or

parts of them, drowned, but also the general erosion of the remaining land is diminished. Such a drowned valley becomes an *estuary*, and the former tributaries of the main stream, now forced to enter tidewater by separate courses, are said to be *dismembered*. The recently sunken coast of Maine is a fine illustration of many drowned-river valleys. The drowned valleys of the lower Susquehanna River (Chesapeake Bay) and of the lower Hudson River are also good examples of such estuaries. The submerged valley of the Hudson has been definitely traced across the sea bottom for fully 100 miles out from New York City (Fig. 39).

Other causes of interrupted cycles. It should not be presumed, from the foregoing statements, that interruptions of the normal cycle of erosion are brought about only by changes of level of the land. Thus the whole northeastern portion of the United States from Minnesota and Iowa to the New England coast was in a topographic condition varying from maturity to early old age just before the great Ice Age. Then, during and after the recession of the vast glacier (Fig. 249) from the region, extensive deposits of glacial and post-glacial rock débris were left more or less irregularly strewn over much of the surface, giving rise to many low hills, lake basins, and altered drainage courses, the last of which have not uncommonly developed gorges and waterfalls. Thus many distinct features of a youthful topography are, as a result of glaciation, superimposed upon a large land area which was otherwise well along in its erosional history.

Extensive outpourings of lava may profoundly interrupt the normal cycle of erosion of a region as is so well exemplified on a grand scale in the Columbia Plateau stretching from the Yellowstone Park region westward across Idaho and into eastern Oregon and Washington. The old erosion surface with its stream systems was almost completely buried under the thick accumulations of lava, and new stream courses have been established upon the newer surface of the lava fields.

Cycle of Erosion in Deserts

Streams of Arid Regions. The cycle of erosion under arid climate conditions shows characteristic differences from the normal cycle in humid regions. Rainfall and, therefore, vegetation are scant. An arid-climate characteristic is that the rain which does fall is likely to be concentrated in a few downpours, each of very short duration, in the course of a year or possibly several years. Large trunk streams seldom

if ever develop. A few only of the stronger streams flow the year round, and most of their tributaries contain water only during, and shortly after, the rare periods of rain. The valleys of the perennial streams are, therefore, nearly always much larger in proportion to the average volume of water flowing through them than those of humid regions. It is commonly the case that parts of a stream bed contain water, and other parts do not. It is also characteristic of arid regions that most of the valleys contain no water most of the time. During the rare, short periods of heavy rainfall (sometimes "cloud-bursts"), water at high speeds and in large volumes rushes through the valleys, but, within a few hours (or days at most), the stream channels (except the few perennial ones) again become dry.

Stages of a Typical Desert Cycle of Erosion. The arid cycle of erosion is so much influenced by the nature of the original topography of the region that the order of events in a cycle varies considerably. Our present purpose is to discuss only the more general principles as they would be illustrated in a rather typical region of varied topography, in the form of a great general basin, within which there are numerous mountains and minor basins, with no stream outlet to the sea, and undisturbed by changes of level of the land during the cycle. The so-called Great Basin of Nevada, Utah, and California may be kept in mind as an example now in the midst of a typical desert cycle. It varies in altitude from somewhat below sea level to more than two miles above sea level, with many mountain ranges and ridges rising conspicuously above the general level of the great irregular floor which is separated into many more or less local closed basins. The present cycle of erosion of the Great Basin began with the breaking up of the region into many fault-block mountains of various sizes and shapes, but most of them roughly parallel, tilted blocks. Beginning with an initial topography of this kind, accompanying Fig. 205 gives a good idea of several stages in the cycle of erosion under desert conditions.

A great, typical desert region like that just described has, in *infancy* of topographic development, consequent drainage courses established upon the initial fault-block surfaces. These streams, which are seldom active except during and just after heavy rains, do not become tributary to a perennial trunk river draining the whole great arid basin or a large part of it, but they flow down upon the floors of the various local basins where they may sink away or evaporate; or in a few cases enter permanent lakes; or more often form temporary, or *playa*, lakes. Most

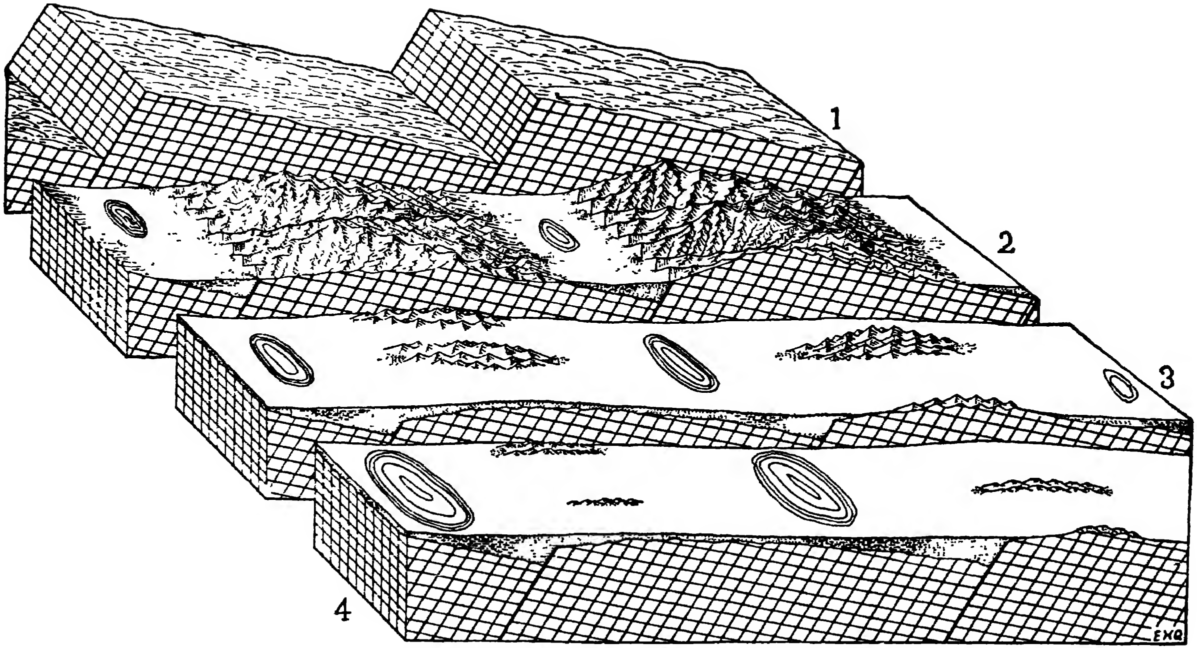

FIG. 205. Block diagrams illustrating a desert cycle of erosion. 1, represents fault-block mountains as they would appear if unaffected by erosion; 2, youthful stage; 3, mature stage; and 4, old-age stage. Important features are explained in the text. Cross-lines indicate bedrock; stippled portions, sediments; and parallel-lined areas, playas.

of the streams are, therefore, mere fragments of what, in a humid region, would be a river system with its trunk river and numerous tributaries. Number 1 of Fig. 205 shows a series of fault blocks in a desert region as they would appear if unaffected by erosion. Slight erosion of these blocks would bring them to the infancy stage, intermediate between numbers 1 and 2 of Fig. 205.

Fig. 206. Youthful stage of erosion of a desert mountain of soft, strongly tilted strata, showing very rugged relief. About 30 miles southwest of Baker, California.

"In the *youth* of the cycle the highlands are slowly eroded, and deposition takes place on the slopes and floor of each basin, diminishing the relief and raising the local base level, a strong contrast to the corresponding stage of the normal cycle in which relief is increased by the excavation of stream valleys. Even in arid regions, however, valleys are cut on the highland slopes, while the basin floor is made nearly level by deposition. This stage is exemplified by the Great Basin and its mountains. Water is the chief agent of erosion and deposition during the period of youth, but the wind is also important in eroding the bare rocks, and in distributing the finer waste, part of which it carries outside of the arid region altogether. Extremely slow as this process of complete removal of the finer débris by the wind undoubtedly is, yet it is the only agency which actually lowers the average altitude of the region,

for no water flows out of the area we are considering" (W. B. Scott).

In regard to the Great Basin it would be better to say that the stage now represented varies from earlier youth to earlier old age. This is because, in this vast region, not all the fault blocks are of just the same age, and also because some of the initial block mountains were much larger than others. Certain parts of the Great Basin are, therefore, farther along in the cycle than others.

FIG. 207. An aerial view showing coalescing alluvial fans forming a bajada on the eastern side of Emigrant Wash in Death Valley, California. (Photo furnished by National Park Service.)

In the youthful stage (Fig. 205, No. 2) alluvial fans are prominently developed along the bases of the mountains, and they extend well out into the desert basins. When a series of these fans run together along the base of a mountain, the whole deposit is called a *bajada* (Fig. 207).

The finer sediment, carried out of the mountains and down the alluvial slopes at times of heavy rain, is moved by part of the flood water to the bottom of the basin where the sediment, often very fine-grained and claylike, is deposited as a broad, thin sheet under water covering a playa. "Nearly every desert basin has its central playa—the

bed of an ephemeral lake. Indeed it is certain that nearly all the playas have in former times been covered by more permanent lakes. They are so level that a sheet of water less than a foot deep may completely cover a plain five miles in diameter. It is doubtful whether any topographic feature of the earth's surface equals the playa in flatness. The typical large playas are underlain by level beds of clay alternating with salt and gypsum. The plain is perfectly bare of vegetation, and presents a glaring buff or white surface" (Blackwelder).

FIG. 208. A desert basin (bolson) 15 miles long with a playa (white area) in the distance. The sediment-covered floor of the basin is a mile above sea level. This basin has no outlet. Deep Spring Valley, California.

A whole desert basin lying between two ranges and being filled with sediment is known as a *bolson* (Fig. 208).

Maturity of the region is reached when the highlands are deeply eroded, and enough of the resulting sediments has been carried down and deposited on the floors of the original separate basins to cause them to coalesce. "As the coalescence of basins and the integration of stream systems progress, the changes of the local base levels will be fewer and slower, and the obliteration of the uplands, the development of graded piedmont slopes, and the aggradation of the chief basins will be more and more extensive" (W. M. Davis).

During the mature stage or even somewhat earlier, broad, gently

inclined, comparatively smooth bedrock surfaces, called *pediments*, are formed around or along the bases of the dwindling mountain masses. They are caused by vigorous erosive action of streams sweeping from side to side, from time to time, as they emerge from the mountains. Pediments are shown on the right in numbers 3 and 4 of Fig. 205.

During maturity the erosive action of the wind becomes relatively

Fig. 209. Looking across Death Valley, California, from near Dante's View (altitude, over 5000 feet) to the Panamint Mountains climaxed by Telescope Peak (altitude 11,045 feet) 21 miles distant. The large white area at the lower right, covered with salt, is about 270 feet below sea level. Note the two large alluvial cones at and near the base of the mountain. (Telephotograph by Dick Freeman, Los Angeles.)

more important not only because rainfall is less on the lowered highlands, but also because the swiftness and hence erosive power of the streams are much reduced on account of the lower relief.

As *old age* is approached the original highlands are cut down lower and lower, and the now very largely coalesced, local basins are built up more and more until the whole region becomes a wide, nearly flat expanse with broad, gently sloping pediments, surrounded by scattering, small, subdued, bare-rock hills, merging into still broader plains of

deposition (filled basins). Such a combination plain may lie far above sea level. The bare-rock hills which rise above the general level, in the form of small residual masses of the original mountains, correspond to monadnocks of the normal cycle of erosion, whereas the pediments represent more or less extensively planed-down parts of the original mountains. The wind is the most active agent of erosion during old age. By

FIG. 210. An aerial view of rugged desert-mountain spurs being buried under alluvial fan material with island-like portions (*inselbergs*) not yet covered. South of San Felipe, Lower California. (© Spence Air Photos.)

its corrasive power it helps to cut down the exposed bedrock. By its transporting power the wind causes much shifting about of the finer sediments within the general arid region, and it also carries some out of it. The latter process is the only one by which the general level of the old-age surface is reduced.

Given a much longer time still, during which there are no important modifying diastrophic effects and a very dry climate, the wind may carry enough sediment (as dust) out of the arid region to lay bare wide portions of the sediment-covered bedrock so that wind-swept rock sur-

faces become very extensive. This stage, later than number 4 of Fig. 205, is seldom reached.

Stream Deflection and Adjustment

Causes of Deflection. There is a strong tendency for streams, especially the larger ones whose valleys have not reached old age, to swerve but little from courses once well established. Smaller streams and, much more rarely, larger ones may, however, be notably deflected from established courses. There are many causes of stream deflection, some of which will now be briefly considered.

The *rotation of the earth* on its axis is an appreciable general cause of deflection of streams. According to Ferrel's law, "if a body moves in any direction on the earth's surface, there is a deflecting force arising from the earth's rotation which deflects it to the right in the northern hemisphere, but to the left in the southern hemisphere." Streams respond to this law and, therefore, tend to swing against and erode their right banks more than their left, thus causing a general shift of stream courses to the right. Streams flowing north accordingly tend to cut their east banks more than their west banks, and those flowing south their west banks more than their east banks. Differences in resistance or structure of the rocks which are being eroded may more than offset this moderate deflective influence. The influence is not only greatest on north- or south-flowing streams, but also it increases with distance from the equator as latitude differences in speed of rotation increase. The streams which flow south across the sloping plains of the southern half of Long Island seem to illustrate the principle. Streams are there cutting shallow valleys into loose, nearly homogeneous sediments with steep banks mostly on the right (west) due to more active erosion there. It has been estimated that the tendency of the Mississippi River to swing against its right (west) bank is about 9 per cent greater than toward its left bank.

A very simple case of stream deflection may be caused by a *lava flow*. Thus, a stream of lava flowed into and across the channel of the Little Colorado River of Arizona, filling it to such an extent that the river has been forced to find a new course around the lava.

In a manner very similar to that of a lava flow, a *glacier* may flow into a valley, forcing the stream over to one side of the valley. Some of the great glaciers of Alaska, as for example certain ones in the Copper River region, provide excellent illustrations.

The great glacier of the Ice Age filled a portion of the valley of the Columbia River of central Washington, forcing the mighty river to find a new course for many miles, along which it eroded a canyon called the Grand Coulee. On the melting of the ice, the river returned to its former channel.

The Missouri River, which was forced by the great glacier to shift to a new course many miles farther west in South Dakota, has kept to the new course since the disappearance of the ice.

Many streams have been notably deflected from their former courses as a result of the blockading of their valleys by accumulations of *glacial deposits* such as moraines (Chapter IX). Thus the combined Monongahela and Allegheny Rivers of western Pennsylvania flowed northward into the Lake Erie basin before the Ice Age. The old valley was so much filled with glacial débris, which was deposited at the border of the great glacier, that the combined Monongahela and Allegheny waters were forced southward into the Ohio Valley.

Another excellent illustration is the lower Sacandaga River of New York which formerly flowed southward into the Mohawk Valley but now flowes northeastward into the upper Hudson River. This deflection was caused by heavy accumulations of glacial débris across the old valley near Gloversville during the Ice Age.

Wind-blown materials often cause shifting of stream courses, as for example the Grand Calumet River which once flowed into Lake Michigan in Indiana. Drifting dune sand so blocked its mouth as to reverse the course of the stream which now empties into the lake near Chicago, about 18 miles from its former mouth.

It often happens that rapid building of an *alluvial cone* or *fan* on the floor of a valley by a tributary forces the main stream of the valley to flow around the edge of the fan or cone, as illustrated by the Illinois River near Peoria. Landslides also often produce similar effects.

Stream deflection on *deltas,* as well as on alluvial fans, is common, as we have already shown in the explanation of distributaries. The Colorado River, which has built a great delta across the upper portion of the Gulf of California, has a usual course into the Gulf, but sometimes it has been deflected down the northern slope of the delta and into Salton Sink.

Tributary streams, on entering graded or aggrading valleys, are often so deflected by *natural levees* along the main streams that they flow considerable distances roughly parallel to the main streams before joining them. The junction is usually effected where the main stream swings far over to the tributary's side of the valley. The Yazoo River of Mississippi is thus deflected for 200 miles before joining the Mississippi at Vicksburg.

Some interesting cases of stream deflection have been caused by so-called "jams" or "rafts" of trees and logs which have floated downstream. Rafts of this kind have grown to a remarkable extent in the lower course of the Red River in Louisiana, this river below Alexandria having been thus diverted for many miles to a new course nearly at right angles to its old course.

Earth-crust movements (*diastrophism*) may cause streams to change

their courses more or less. Thus at the time of the great Assam earthquake in India in 1897, movement of earth blocks along a fault fracture parallel to a meandering river caused many local changes in the stream course. An example on a larger scale, observed by the writer, is a mountain ridge which has risen about 1000 feet, as a fault-block, athwart the path of a stream. Across the top of this ridge, facing Deep Spring Valley in middle-eastern California, an abandoned river bed, strewn with waterworn boulders, proves that the ridge rose too fast for the stream to maintain that part of its course.

We have already shown how streams gradually sweep in broad, meandering curves back and forth from one side of a valley floor to the other when they are in a graded or nearly graded condition, and how streams abandon such *meanders* by cutting across their narrow necks (Fig. 159). The great flood plain of the lower 500 miles of the Mississippi River furnishes many fine examples.

Finally, a very common kind of stream deflection results from the capture of part of one stream by another and the diversion of the water of the one into the other. This interesting and important process of stream development is separately discussed under the next heading.

Stream Capture. *General principles.* During the erosional history of a region it often happens that certain streams steal parts (or all) of other streams by a process known as *stream capture* or *piracy*. The general principle involved is that a stream which finds various conditions for valley development (erosion) more favorable than a near-by stream may, by headward extension of itself or one of its branches, tap and divert into itself part (or all) of the stream whose erosional conditions are less favorable. A stream whose upper waters have been captured is said to be *beheaded*. Through the process of stream capture there is a strong tendency for many streams to leave the harder, or more resistant, rocks and develop courses in softer, or less resistant, rocks, that is, they tend to adjust their courses to the character and structure of the various rock formations of a region. This is known as *structural adjustment* of streams.

Examples of stream capture. Some of the more common principles of stream capture may be made clear by explanation of a few definite cases. Thus, two streams flowing roughly parallel to one another may each develop a tributary reaching out toward the other as shown by Fig. 211. Because one of these streams is more active and has cut its valley deeper, its tributary also cuts down and works headward faster than the tributary of the other stream. The head of the more active

tributary finally reaches the less active tributary and carries off its upper waters. This is a common method of stream capture in many regions.

Where two streams follow approximately parallel courses (one higher than the other) lateral erosion of one or both may at some place completely remove the divide which separates them, causing the stream at the higher level to drain into the lower-level one.

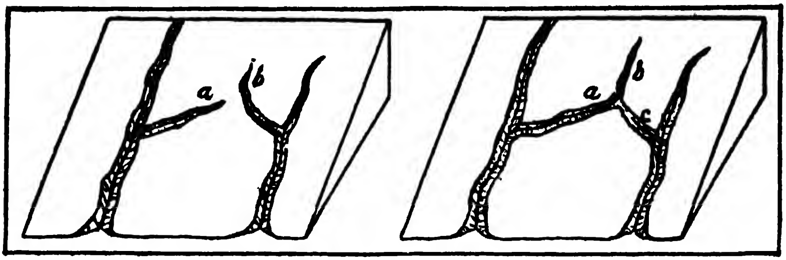

FIG. 211. Diagrams illustrating a simple case of stream capture by headward growth of a tributary. (Modified after Salisbury.)

The principle of stream capture by shifting of divides is very clearly illustrated in the Catskill Mountains of New York. As shown in principle by Fig. 212, two small, very swift streams (*a* and *b*) flow down

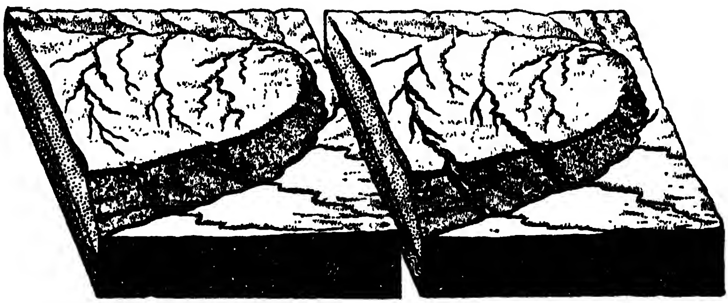

FIG. 212. Diagrams illustrating the principle of stream capture in the Catskill Mountains, New York. (From Tarr and Martin's *College Physiography* by permission of the Macmillan Company.)

the steep front of the mountains, while three much slower streams, tributary to *c*, flow down the more gradual slope on the opposite side of the divide. The short, swift streams have extended their valleys headward so rapidly that several branches of two of the tributaries to *c* have, one after another, been diverted into the shorter streams. In this

case the drainage direction of the diverted streams has been, in the main, reversed, and the original divide has been notably shifted.

The capture of the upper part of Beaverdam Creek by the Shenandoah River of Virginia is a well-known case of stream piracy. As shown by Fig. 213, the Shenandoah developed as a tributary of the Potomac in an early stage of the erosion of the newly uplifted region. Both the Potomac River and Beaverdam Creek cut gorges through the

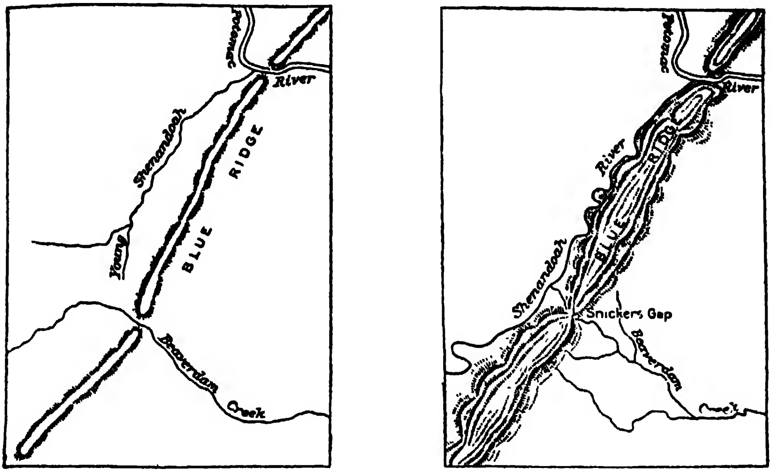

FIG. 213. Sketch maps showing how the upper waters of Beaverdam Creek were captured by the Shenandoah River. (After B. Willis, U. S. Geological Survey.)

hard rock of Blue Ridge, but the former deepened its valley much faster. The young Shenandoah was, therefore, enabled to extend its course southward by headward erosion, and finally tapped and diverted into itself the upper part of the higher level Beaverdam Creek. The abandoned channel of the creek across the Blue Ridge is still plainly preserved.

The short, swift rivers which flow down the western side of the Andes Mountains of Chile have captured the source streams of many of the longer, slower rivers which flow down the eastern side of the mountains and across Argentina.

Antecedent Streams. A type of river of special interest is one which during, and for a time at least after, disturbance (by diastrophism) of its drainage area maintains the course it had before the disturbance

began. Such a stream is said to be *antecedent* because its course was established before the land across which it flows was disturbed by earth-crust movement. The simplest case is that of a revived river resulting from rejuvenation of a region by uplift without much change in the general direction of slope of the land. Thus the rivers of central and western New York have, as already explained, cut valleys in a rather uniformly upraised peneplain. Since such antecedent rivers merely renew down-cutting along their old courses, it is, perhaps, just as well to call them simply revived rivers.

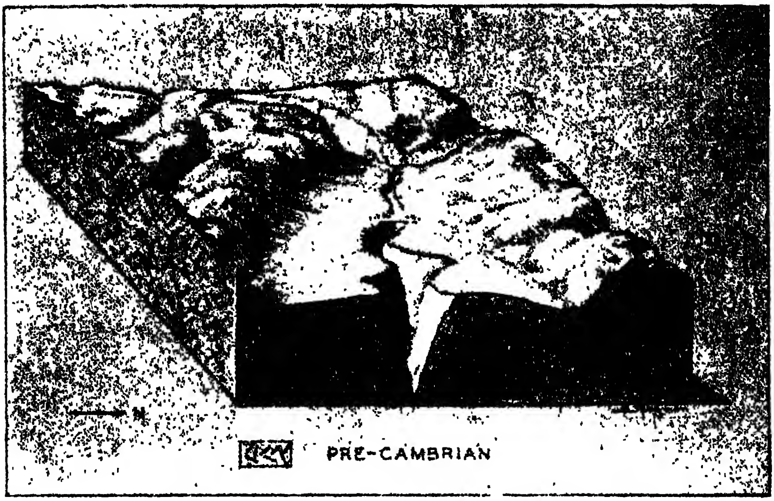

Fig. 214. Block diagram showing the antecedent nature of the Arkansas River where it has cut the famous Royal Gorge in hard pre-Cambrian gneiss and through the rising Front Range of Colorado. (After Atwood and Atwood, Bul. Geol. Soc. Amer., Vol. 49, No. 6, 1938.)

A remarkable type of antecedent river is one which has kept its course through a rising barrier, even a mountain range (Fig. 214). Thus the Arkansas River has maintained its course right across the slowly rising Front Range of Colorado by cutting a narrow canyon more than a thousand feet deep in hard rock (pre-Cambrian gneiss). This is the famous Royal Gorge illustrated by Figs. 178 and 214. A railroad has with great difficulty been built through the bottom of the canyon.

As the Wasatch Range of Utah slowly rose (in recent geologic

time) across the path of the Ogden River, the river maintained its course by cutting a deep canyon.

The Indus and Brahmaputra Rivers of northern India are believed to be antecedent, for they cut great canyons through a main range of the Himalayas and then flow into the Indian Ocean.

Superimposed Streams. An old land mass with characteristic topography, rock character, and structure may be buried under later rock formations of very different kinds and arrangement. The newer, overlying accumulations may be sheets of lava, volcanic ash-beds, glacial deposits, lake deposits, or marine strata. The surface of the newer formation may be utterly different from that of the older, underlying formation.

A simple case to consider is a series of horizontal strata, with a smooth surface, resting on top of a rugged surface of igneous and irregularly tilted metamorphic rocks. It not uncommonly happens that a stream, whose course has been determined upon the newer surface, cuts through the overlying rocks and into the underlying rocks, maintaining its course irrespective of the surface, character, and structure of the underlying rocks. Such a stream is said to be *superimposed* or *inherited.* A fine example is the Colorado River in the Grand Canyon of Arizona where the river has cut through a thickness of several thousand feet of nearly horizontal strata and into a very ancient, worn-down, buried mountain area consisting of a complex arrangement of hard igneous and metamorphic rocks. In the strata the canyon is wide and terraced, but in the hard, underlying rocks a deep, narrow, steep-walled gorge has been (and is being) cut by the river.

Part of a plateau in western Colorado consists of nearly horizontal beds of volcanic rocks resting upon tilted strata which in turn cover an old, very uneven-surfaced, mountainous mass of hard, ancient (pre-Cambrian) gneiss. The Gunnison River, starting on the plateau surface, cut its way through the volcanic and stratified rocks and 1700 feet into (and across) the uneven-surfaced gneiss, meantime maintaining its original course. Thus the narrow, steep-walled Black Canyon of the Gunnison was produced (Fig. 215).

Where, through erosion, the overlying rock mantle has been completely removed from the underlying rocks of different structure, the superimposed streams may have courses very strikingly out of harmony with the structure of the formerly buried rocks. Thus in the Lake

District of England a system of streams with a distinctly radial arrangement has been superimposed upon a once buried body of rocks with a northeast-southwest trend. This is a fine example of a ***superimposed,*** or ***inherited drainage, system.***

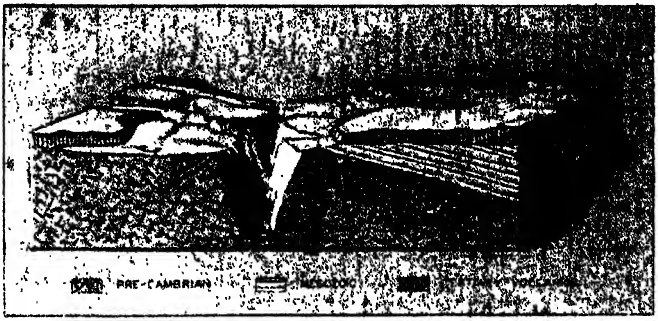

FIG. 215. Block diagram showing the superimposed nature of the Gunnison River, Colorado, at the Black Canyon of the Gunnison. The river has cut through, and partly removed, a cover of Mesozoic strata and Tertiary volcanic rocks and into an underlying mountainous mass of ancient (pre-Cambrian) hard gneiss. (After Atwood and Atwood, Bul. Geol. Soc. Amer., Vol. 49, No. 6, 1938.)

SPECIAL EFFECTS OF STREAM WORK

Canyons and Gorges. The principles of ordinary valley development through the agency of stream erosion have already been discussed. We shall now briefly consider the special type of valley which is exceptionally deep in proportion to width. Such relatively deep, narrow, steep-walled valleys are called gorges (e.g., Niagara Gorge), chasms (e.g., Ausable Chasm of New York), dells (e.g., Dells of the Wisconsin), glens, (e.g., Watkins Glen of New York), or canyons (e.g., the Grand Canyon of Arizona). They are among the most spectacular scenic features enjoyed in travel. The term "canyon" is generally applied to a large gorge or chasm.

Factors favoring canyon development. Factors particularly favorable to the development of canyons and gorges are (a) altitude high above base level, (b) rapid down-cutting by streams, (c) rock formations resistant enough to maintain steep slopes or cliffs when they are cut into, (d) an arid climate which is usually more favorable than a moist one because certain weathering agents which cause valley widen-

ing are less effective under dry climate conditions. In the development of a gorge or canyon the down-cutting action of a stream proceeds so rapidly that the agents of valley widening do not have time to reduce notably the steepness of the valley sides.

Zion Canyon. A remarkable example of a deep, very narrow canyon is the northern portion (so-called "Narrows") of Zion Canyon, Utah, where a very swift, sediment-laden stream under semi-arid conditions has cut its way down into moderately hard rock (sandstone) so fast as to develop a gorge over 1500 feet deep, 20 to 40 feet wide at the bottom, and a few hundred feet (or less) wide across the top.

The main part of Zion Canyon, some 12 miles in length, varies from 2000 to nearly 4000 feet deep. It is bounded by precipitous walls of red, horizontal beds of massive sandstone overlain with light gray sandstone. The canyon gradually gets wider toward its mouth. Fig. 216 gives a fine idea of the canyon about 10 miles from its mouth.

The great depth of Zion Canyon is entirely the result of the down-cutting (erosive) action of the North Fork of the Virgin River. The canyon has been cut into the Colorado Plateau near its western border. Great rejuvenation of this part of the plateau by uplift on the east side of the Hurricane fault (Figs. 101 and 102) in the present (Quaternary) period has caused great revival of activity of both the Virgin River and its North Fork. Down-cutting by North Fork has, therefore, been so fast that valley-widening processes have not had time to change the profile of the canyon to a V-shape.

Kings River Canyon. A canyon remarkable for its combination of narrowness and depth is the Kings River Canyon of the Sierra Nevada Range of central California. This steep-sided, V-shaped canyon has been carved out of solid granite by the erosive action of the river, aided by relatively little weathering, to the amazing depth of 6900 feet. Profound uplift and tilting of the Sierra earth block (Fig. 203) in recent geologic time; volume and swiftness of the water; hardness of the rock; and a liberal supply of grinding tools are the conditions which have favored the development of this canyon.

Yellowstone Canyon. The Yellowstone River of Yellowstone National Park has cut a narrow, steep-sided canyon (Fig. 194) over 1000 feet deep and 15 miles long into a high plateau which was built up by outpourings of vast sheets of lava in recent geological time.

The Royal Gorge. The famous Royal Gorge of Colorado has been (and is being) cut through the recently uplifted Front Range of the

Rocky Mountains. It is remarkably narrow with almost vertical walls rising to a height of 1100 feet (Fig. 178).

Grand Canyon of Arizona. Greatest of all canyons, not only of North America but also of the world, is the Grand Canyon of the Colorado River in Arizona. Its general dimensions are: length, over

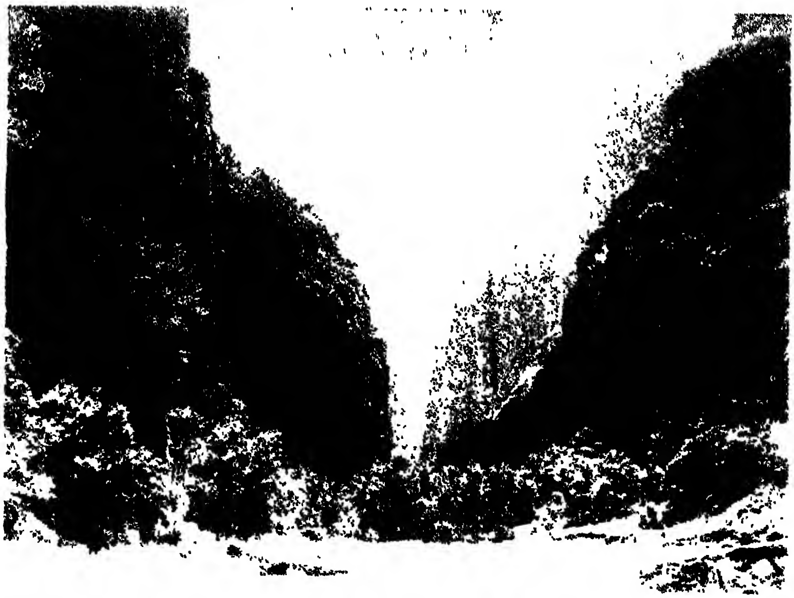

FIG. 216. A view of Zion Canyon, Utah, about 10 miles from its mouth. The precipitous walls of red sandstone below and light gray sandstone above are here about one-half of a mile high. (Photo by W. T. Lee, U. S. Geological Survey.)

200 miles; depth from 4000 to 6000 feet; and width from 7 to 15 miles (Fig. 217). This mighty gash in the earth's crust has been excavated wholly by the Colorado River and some of its shorter tributaries, aided by weathering. Some of the conditions exceptionally favorable to this canyon development have been and are: (1) The recent great uplift of the region, providing a thickness of many thousands of feet of rocks to be cut through by the river before reaching grade; (2) the large, very swift river; (3) the abundance of rock fragments constantly carried by the river, thus providing for continually aggressive corrasive action; (4) rock formations hard enough and so arranged that most of them

are capable of standing in cliffs or steep slopes; and (5) the arid climate which causes valley widening to be relatively slow.

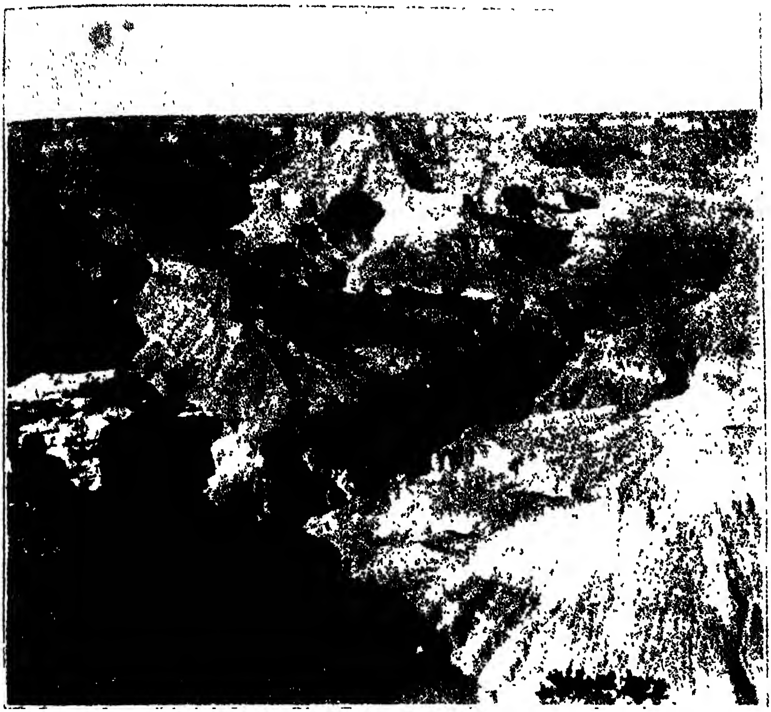

FIG. 217. A view across the world's greatest canyon. Grand Canyon of Arizona. (Photo by courtesy of U. S. Reclamation Service.)

The canyon has been carved out during the present period of geological time, and all the solid rock which once filled the space now occupied by the canyon has been carried away by the Colorado River.

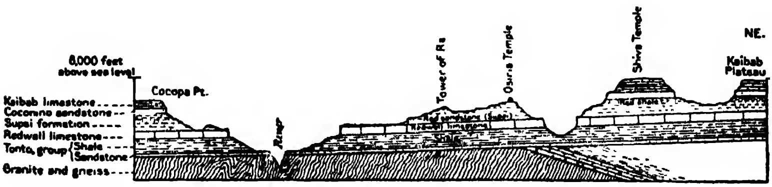

FIG. 218. Structure section across the Grand Canyon of Arizona showing the shape of the canyon, and the kinds and structural relations of the rocks. (After N. H. Darton.)

Much of the sediment has been deposited in the form of a great fan-shaped delta near the mouth of the river, while the rest has gone into the Gulf of California. The Grand Canyon is still being widened and deepened because the swift, active Colorado River in the bottom of the canyon is still 2000 to 3000 feet above sea level and far from graded. The maze of side canyons is due to erosion by tributaries to the main river. The numerous buttes and mesas, often of mountainous proportions, rising within the great canyon are erosional remnants which have not been reduced as fast as the rest of the rock mass by erosion.

The rocks of the main or broader part of the canyon are all strata of Paleozoic age. They form a vast pile of nearly horizontal layers reaching a total thickness of nearly 4000 feet as shown in Fig. 218. These strata include sandstone, shale, and limestone. The outcropping edges of the formations, which are colored light gray, red, and greenish gray, produce the striking color bands so clearly traceable for many miles within the canyon. Outcropping edges of the more resistant formations are in the form of great and small cliffs, while the weaker formations yield steep slopes, often talus covered.

In the narrow, V-shaped inner gorge, through which the river flows in the depths of the canyon, the rocks are mainly dark colored schist, granite, and gneiss of Archeozoic age.

At the beginning of the present period of geological time the Colorado Plateau region was an old-age surface on which the Colorado River flowed with a meandering course not far above sea level. Rejuvenation of this old surface by uplift to an altitude of 7000 to 8000 feet in the canyon region greatly revived the erosive activity of the river which has carved the mighty chasm out of the uplifted (Colorado Plateau) region.

Canyons modified by glaciers. Many deep canyons in the mountains of the western United States, western Canada, and southern Alaska are not wholly the work of running water. Such canyons were cut to great depths by streams after which (during the Ice Age) they were occupied for many years by streams of ice called glaciers which deepened and broadened the bottoms and steepened the sides of the canyons. Excellent examples are the Swiftcurrent Canyon of Glacier Park, Montana (Fig. 262), and the famous Yosemite Valley (or Canyon) of California (Fig. 263).

Examples of gorges. There are numerous examples of gorges and small canyons in the eastern United States, such as the Ausable Chasm near Plattsburg, New York, and Watkins Glen of southern New York,

each of which is from 20 to 50 feet wide, and 100 to 200 feet deep; the Flume in the White Mountains of New Hampshire; and the gorges of Tullulah River in Georgia, and the French Broad River in North Carolina, each of which is many hundreds of feet deep.

Narrows and Gaps. River narrows and water gaps are in reality only special types of gorges or canyons. When, during the process of its valley development, a stream takes its course across a belt or irregular mass of rock which is relatively more resistant, the valley is there carved out less widely and rapidly than in the weaker rocks just upstream and downstream from the harder rock. Local contractions of river valleys, formed under such conditions are called *narrows* or, if they are very short, *water gaps*. Rapids, cascades, and low waterfalls are common in river narrows and gaps. The more resistant rock athwart the channel locally slows up the process of down-cutting, and so there is a tendency for a "temporary base-level-of-erosion" to be established for a greater or less distance upstream from the harder rock.

A few of the many well-known examples of river narrows and water gaps will be cited. The Mohawk River at Little Falls, New York, flows for nearly two miles through a narrow, steep-sided gorge hundreds of feet deep in hard rocks, while for many miles above and below the narrows the valley has been opened out widely on weak rocks (mostly shales). The lower Hudson River has cut a narrows hundreds of feet deep and 16 miles long through hard granite and related rocks. Near Northampton, Massachusetts, the Connecticut River has eroded a water gap hundreds of feet deep through the Holyoke Range of hard lava, while the broad valley has been opened up by the river in weaker, stratified rocks both above and below the gap (Fig. 198). The famous Delaware Water Gap has been cut by the Delaware River through a tilted formation of hard conglomerate, on either side of which there are relatively weak strata. The Susquehanna River near Harrisburg, Pennsylvania, flows through a succession of typical water gaps where tilted, resistant, rock formations, with intervening, weak rocks extend across the course of the river.

A water gap abandoned by its stream becomes a so-called *wind gap* because of the tendency for the wind to blow with unusual force through the narrow opening in the ridge. Wind gaps very commonly result from stream piracy where a stream flowing through a water gap has its course diverted by a neighboring stream. The principle involved is perfectly illustrated in the vicinity of Harper's Ferry, Virginia (Fig.

213), where the water gap of Beaverdam Creek was converted into a wind gap (called Snicker's Gap) because of the capture of the upper waters of the creek by the Shenandoah River. There are numerous wind gaps in the central and southern Appalachian Mountains similar in origin to Snicker's Gap. Many of them are notches in the tops of the mountain ridges. One of particular interest is Cumberland Gap, on the Kentucky-Virginia line, which is a pass 700 feet deep through the Cumberland Mountain ridge. Several hundred thousand immigrants traveled through this wind gap on their way west in the latter part of the eighteenth century.

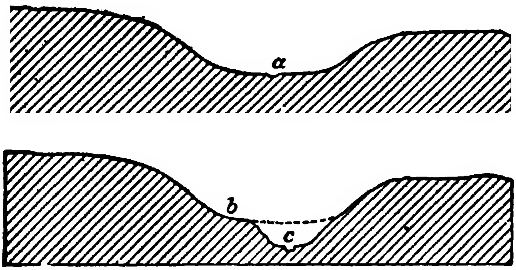

FIG. 219. Diagrams illustrating the development of a rock terrace by a stream. (After U. S. Geological Survey.)

Stream Terraces. Along the sides of a valley there may be benches or nearly flat surfaces with steep fronts facing the stream in the valley and too high to be covered by flood-waters. Two or more of them may be arranged one above another in steplike form on both sides of the valley. Such benches, when formed by the action of the stream, are called *stream terraces*. Two of their most common modes of origin will now be explained.

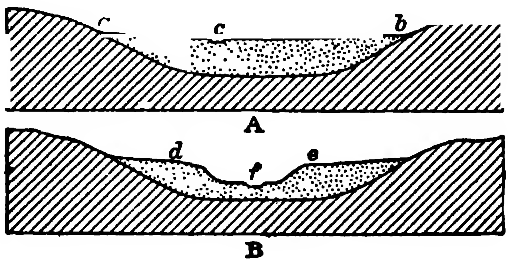

FIG. 220. Diagrams illustrating the development of alluvial terraces. (After U. S. Geological Survey.)

Rock terraces. We have already learned that a stream, on approaching grade in its down-cutting process, begins to widen its valley floor notably by meandering back and forth from one side of the valley to the other. A flood plain of such a stream may be covered with more or less stream-deposited (alluvial) soil. Uplift of the region may then

take place, causing the revived river to cut a young, steep-sided inner valley (or gorge) into the old flood plain. The remnants of the old valley flat, consisting of bedrock covered with some alluvium, constitute one kind of *rock terraces*. An interesting case is illustrated by Fig. 219. After a flat is developed in the bottom of the newer valley, another uplift would inaugurate the development of terraces at a still lower level. Rock terraces not uncommonly develop during the down-cutting of a valley where resistant layers of horizontal or nearly horizontal rocks are worn back on the valley sides less rapidly than weaker layers just above them. A wonderful succession of such rock terraces occurs on a magnificent scale in the Grand Canyon of Arizona, giving rise to what may be called *step topography* (Fig. 217).

FIG. 221. Stream-cut terraces in Fraser River valley near Lillooet, British Columbia. (Photo by Calvin.)

Alluvial terraces. If, for any reason, a valley becomes partly filled with alluvial sediment, and then the stream in the valley has its erosive activity notably revived by either decreased load or uplift of the land, so-called *alluvial terraces* will develop. Rapid down-cutting by the stream may result in only one terrace level. Often, however, the stream cuts down into the alluvial filling slowly enough to allow the development of meanders. The stream then cuts laterally into the alluvium first on one side of the valley and then on the other, in each case leaving a flat with a steep face toward the stream. Swinging back to the opposite side of the valley, this time at a somewhat lower level, a new flat is developed and the earlier (higher level) terrace is partly cut away. By such a process a succession of two or more alluvial terraces may be formed (Fig. 220). Excellent examples occur in the Connecticut Valley of New England and in many other valleys.

Erosional Remnants. *General principles.* During the process of general lowering of lands by erosion, it very commonly happens that certain local portions are not cut down as fast as most of the area, and so are left standing out more or less conspicuously above the general

level of the country as *remnants of erosion.* There are two important causes of such unequal erosion. One is lack of uniformity of character and structure of the rocks of an area, that is, some portions may be either harder, or more resistant, than others, or less subject to weathering because less broken and fissured by joints or faults. Another cause of erosional remnants is the less rapid erosion in the spaces between streams, this being particularly true in relatively level plain or plateau districts. Erosional remnants are variously shaped and named.

Towers and pinnacles. There may be *rock towers, pinnacles,* or *pillars* consisting either of notably harder, isolated masses such as the igneous rock of Devil's Tower, Wyoming, or of the cores of volcanoes (volcanic necks) in Arizona (Fig. 133), or of isolated joint blocks of essentially homogeneous material such as the Cathedral Spires in the Garden of the Gods, Colorado (Fig. 158), or the pinnacles and pillars of lava near Vantage, Washington (Fig. 163), or the remarkable maze of towers and pinnacles in Bryce Canyon, Utah (Fig. 224). In the case of Bryce Canyon the active factors of rain wash and general erosion, aided by the passive factors of stratification, variations in resistance of beds, and vertical jointing, have produced these remarkable forms which are also beautifully colored.

In this connection mention may be made of certain *pedestal rocks* of special interest recently observed by the author in northern Arizona. Fig. 225 shows some of them. These large, angular boulders rolled from a near-by cliff down upon a bed of very soft shale. Heavy rains, involving sheet erosion and numerous tiny rivulets, have washed away most of the shale to a depth of several feet, but the boulders protected the shale under them from the rain, leaving these portions of it in the form of pedestals supporting the boulders often in remarkably balanced positions. Pedestal rocks also form in other ways, as for example by unequal weathering (Fig. 162).

Mesas. If the rocks are in horizontal layers or nearly so, and some are harder than others, flat-topped hills or small mountains, called *mesas* (pronounced "maysas"), often become erosional remnants. In such cases the flat surfaces are determined by harder layers. Similar isolated masses without flat tops are called *buttes* (pronounced "bewts"). Mesas and buttes are common and typical in many portions of the high, arid to semi-arid plains and plateaus of the western United States, particularly the Colorado Plateaus of parts of Arizona (Fig. 222), New Mexico, Utah, and Colorado. Many mesas, buttes, towers, and pinnacles belong in the category of so-called *outliers,* that is, rem-

nants of more extensive bodies of similar rocks separated from the latter by erosion.

FIG. 222. A mesa capped with a lava-bed. About 14 miles southwest of Peach Springs, Arizona.

FIG. 223. A hogback of Mesozoic strata. Near Fort Wingate, New Mexico. (Photo by Hillers, U. S. Geological Survey.)

Ridges. Where erosion proceeds upon a region of highly inclined to vertical (or folded) rock layers or formations which are alternately hard and soft, the hard belts will, especially during maturity, stand out

in relief in the form of *ridges* because erosion cuts down the weaker (softer) rocks more readily, developing valleys in them. This principle

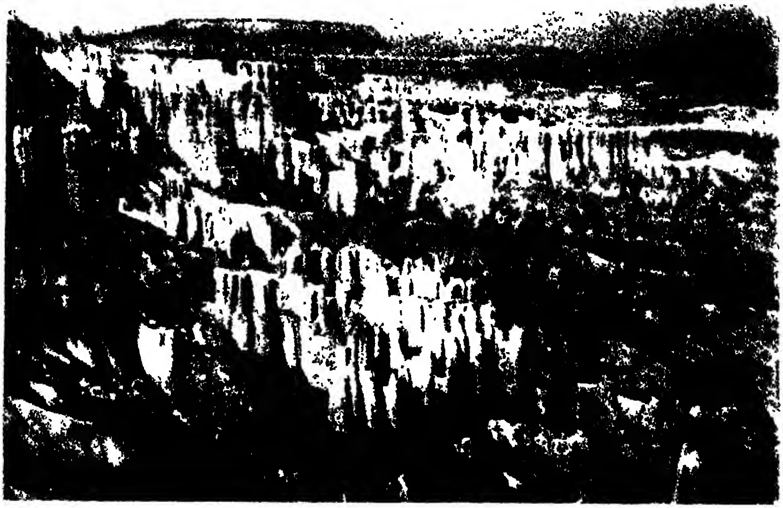

FIG. 224. Bryce Canyon, Utah, showing a maze of towers and pinnacles which have been carved out of highly jointed, rather soft, horizontal strata.

FIG. 225. Large blocks of red sandstone resting upon pedestals of soft shale. Nine miles southwest of Lee's Ferry, Arizona.

is grandly illustrated by the numerous approximately parallel ridges and valleys in the Appalachian region between northeastern Pennsylvania and eastern Tennessee (Fig. 355).

Hogbacks and cuestas. A ***hogback*** is an erosional ridge with a long, relatively gentle slope on one side and a short, steep (or precipitous) slope or face on the other side. Such a ridge develops where rock

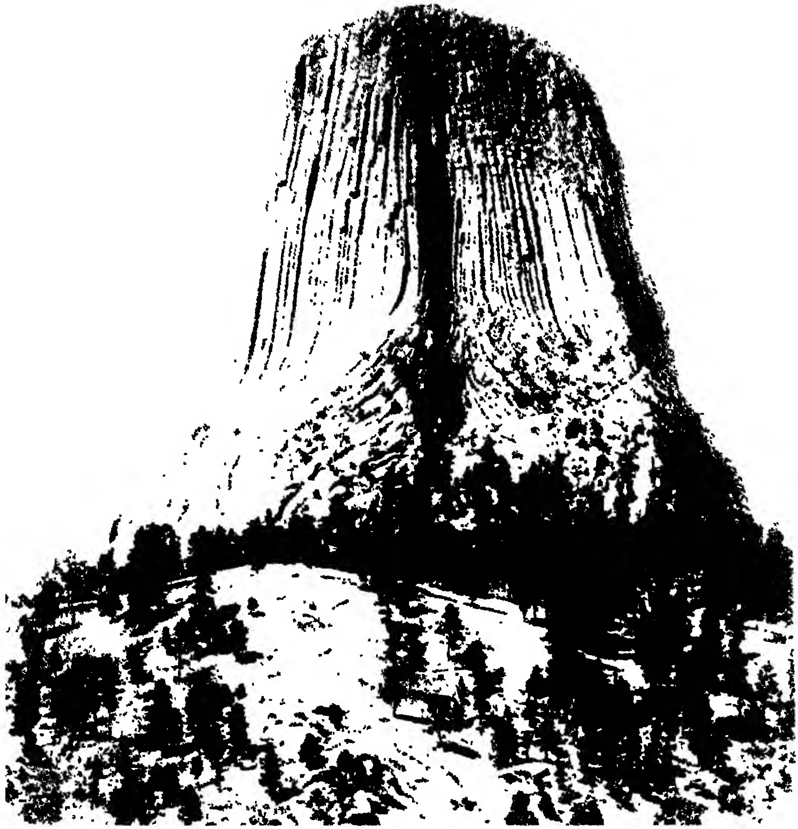

Fig. 226. A great tower-like, erosional remnant 700 feet high. It is a mass of resistant igneous rock left standing after removal of relatively weak strata which once encompassed it. Note the excellent columnar structure. Devils Tower, Wyoming. (Photo by N. H. Darton, U. S. Geological Survey.)

layers (or formations) are moderately tilted with a hard layer lying between soft layers. The long gentle slope is caused by the removal of the weak rock from the top of the hard layer, and the tendency of the weak underlying rock to erode (or weather) faster than the hard, tilted

layer just above it. Hogback ridges and succession of ridges are very typically displayed near the eastern base of the Rocky Mountains in Colorado and also in parts of Arizona and New Mexico (Fig. 223).

A *cuesta* is practically the same in principle as a hogback, but on one side its slope is very long and gentle, while on the other side there is an abrupt slope, or even a cliff. Cuestas are well illustrated in the Atlantic and Gulf Coastal Plains of the United States, and, on a grand scale, in the Colorado Plateau country.

FIG. 227. The great Augusta Natural Bridge in southeastern Utah. (Photo by G. L. Bean, courtesy of the National Park Service.)

Monadnocks. A special kind of erosional remnant is the *monadnock* already described. It represents a residual mass of country rock of any shape which has not been reduced to the general level of the peneplain during a late stage in the erosional history of a region.

Natural bridges. If, during the process of erosion of a region, a stream perforates the neck of one of its rather deeply entrenched (incised) meanders, a *natural bridge* results, as may be readily understood by examination of Fig. 228. The largest natural bridges in the world have originated in this manner, several of them being located in San Juan County, Utah (Fig. 227). Greatest of all is the Rainbow

Bridge which would easily span the dome of the Capitol Building in Washington. It should be clearly understood that natural bridges originate in various other ways than by the action of surface streams.

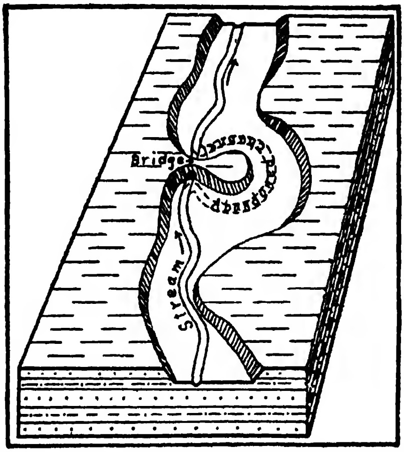

FIG. 228. Diagram illustrating one mode of origin of natural bridges.

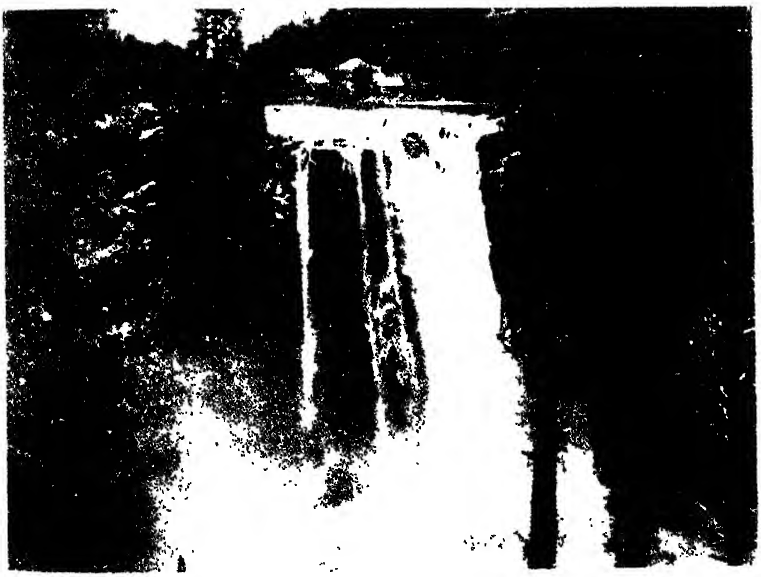

FIG. 229 Snoqualmie Falls in the Cascade Mountains, Washington. Height, 272 feet.

FIG. 230. General view of Niagara Falls and the upper part of Niagara Gorge. American Falls on the left and Canadian Falls on the right. (Photo furnished by Rau Art Studios, Philadelphia.)

Waterfalls. Where a stream rushes over a steep slope in its bed it forms a *rapid*. Where a stream plunges over a vertical or nearly vertical rock face it forms a *waterfall*. Between ordinary rapids and true waterfalls, all gradations exist. Waterfalls are sometimes called *cascades* or *cataracts*. Waterfalls originate in many ways. Our present purpose is to consider only some of the most important principles of waterfall development, with emphasis upon the kinds of falls which owe their existence to the more or less direct erosive action of the streams which themselves form cataracts.

Niagara Falls type. The most common principle is involved in what may be termed the Niagara type of waterfall, so wonderfully illustrated by Niagara Falls (Fig. 230) which is one of the world's greatest cataracts. Its tremendous volume of water, divided into two parts (Canadian Falls and American Falls), plunges about 160 feet. In this type of waterfall, the rock formations lie in an approximately horizontal position with a resistant formation on top of a notably weaker one. At Niagara there is a hard limestone resting upon soft shales in thin layers. The conditions are shown by Fig. 231. Under the influence of weathering and the swirling action of the water, the weaker, underlying rocks are cut away faster than the harder overlying rock, causing the latter to overhang so that blocks of it fall down from time to time, and are mostly carried away by the swift current. The waterfall maintains itself while it retreats upstream and develops a gorge. By this process Niagara gorge, seven miles in length, has been produced since the withdrawal of the great glacier of the Ice Age from the Niagara region—not more than a few tens of thousands of years ago.

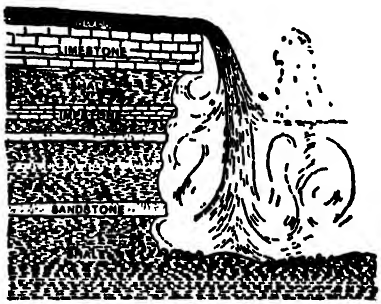

FIG. 231. Structure section at Niagara Falls. (After Gilbert, U. S. Geological Survey.)

At Snoqualmie Falls (272 feet high) in the Cascade Range of Washington, the rocks are of volcanic origin with a more resistant layer on top of a weaker one (Fig. 229). A layer of hard lava rests upon softer lava at the crest of Shoshone Falls where the Snake River of southern Idaho plunges vertically 210 feet.

FIG. 232. Great Falls of the Yellowstone River. Height, 308 feet. Yellowstone National Park. (Photo by Hillers, U. S. Geological Survey.)

Fig. 233. Yosemite Falls, California. Height of upper falls, 1430 feet; lower falls, 320 feet. (Photo by courtesy of the Southern Pacific Lines.)

Yellowstone Falls type. The Yellowstone type of waterfall involves a highly inclined or vertical mass of resistant rock extending across a stream channel, with weaker rock on the downstream (and usually also on the upstream) side of it. At the Great Falls in Yellowstone National Park the river crosses a vertical mass of hard, fresh lava in the midst of other lava which has been much weakened by weathering. This hard rock acts as a barrier, permitting rapid down-cutting immediately on its downstream side, but checking erosion on its upstream side. The river, therefore, plunges 308 feet over the vertical face of the barrier (Fig. 232). Waterfalls of this kind commonly develop also in youthful stages of erosion in regions with highly inclined or vertical rock formations of varying degrees of hardness.

Yosemite Falls type. Another type of waterfall is only indirectly a result of stream erosion. Many of the highest waterfalls of the world belong in this category which we call the Yosemite type on account of the wonderful development of such falls in Yosemite Valley, California. A very active river carved out a deep, steep-sided, V-shaped canyon in the hard granite of the Yosemite region. Then a powerful glacier plowed slowly through the canyon, broadening, and somewhat deepening it, and making its walls precipitous by cutting them back. On the melting of the glacier, various tributaries were forced to enter the main valley by plunging over great granite-cliffs. At Yosemite Falls, a stream plunges the amazing distance of 1430 feet vertically over such a granite cliff, this being the highest true waterfall in the world. The same water, after descending a very steep slope for 800 feet, plunges 320 feet vertically to the floor

FIG. 234. Takkakaw Falls, Yoho Valley, British Columbia. Height, 1200 feet.

of the valley (Fig. 233). Bridalveil Falls in the same valley and of similar origin is 620 feet high. Throughout the mountainous portions of North America and Europe which were occupied by glaciers during the Ice Age, the Yosemite type of waterfall is common. Examples are a fall 1200 feet high (not wholly vertical) in the Yoho Valley of British Columbia (Fig. 234), and one 900 feet high in the Launterbrunnen Valley of Switzerland.

Victoria Falls type. The Victoria Falls of South Africa, probably the greatest in the world, involves a principle opposite to that of the Yellowstone type, that is, a belt of weak rock lies across the course of the river in the midst of hard rock (lava). The Zambezi River, finding

FIG. 235. Detail view of part of Victoria Falls, Zambesi River, South Africa. Full height (not shown) is more than 400 feet. (Photo by A. J. Orner.)

the work of erosion much easier along the belt of weak (highly jointed and fractured) rock, has turned abruptly to follow the weak rock into which it has cut a deep, narrow chasm. The river, which is here over a mile wide, plunges vertically more than 400 feet into the chasm, which is only a few hundred feet wide (Fig. 235).

Trenton Falls type. A common type of waterfall results from the removal of joint blocks of rock. Where the rock in the bed of the stream is traversed by well-developed vertical cracks (so-called *joints*), somewhat loosened blocks of rock may be further freed by weathering and then, one by one, pushed away by the stream. In this manner a vertical face of rock is produced over which the water plunges. As such a fall retreats by removal of joint blocks, a gorge develops. Taughannock Falls (215 feet high), north of Ithaca, New York, and several

falls (one 50 feet high) at Trenton Falls, New York, are good illustrations (Fig. 236).

Other types. Among other types of waterfalls, a few will be mentioned. Thus a stream may plunge over the edge of a roof of a partly collapsed lava tunnel. A good example is Rainbow Falls near Hilo, Hawaii. Lava tunnels are explained in Chapter VI.

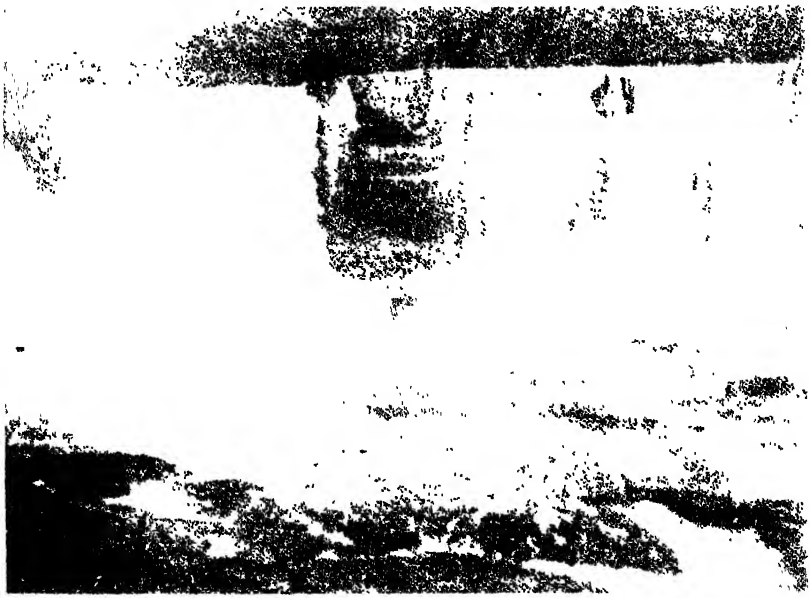

Fig. 236. A joint-face type of waterfall. High Falls (50 feet high) at Trenton Falls, New York.

A stream may, for a short time, form a waterfall by plunging over a recently developed fault scarp.

In some places where sea waves cut into the land, with resultant rapid retreat of sea cliffs, streams headed seaward may be forced to plunge over the cliffs. Good examples are on the Island of Hawaii. Retreat of sea cliffs is explained in Chapter XI.

Potholes. Where rock fragments are given a rapid, swirling motion by an eddy in a swift stream they often wear round or cylindrical excavations, known as *potholes,* in the bedrock of the stream. Such an eddy must of course maintain itself in one place long enough for the grinding action to develop the pothole which may be from a few inches

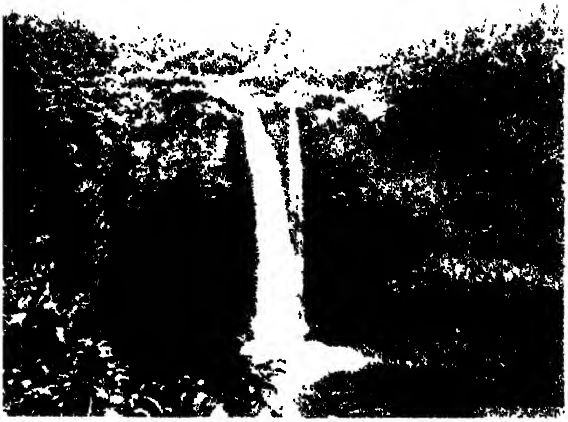

FIG. 237. Rainbow Falls, near Hilo, Hawaii, at time of low water. The rock is lava. Note the lava tunnel.

to 25 feet or more in both diameter and depth. As the grinding materials, consisting of sand, gravel, or even boulders, wear out, new materials are supplied by the stream. Local portions of stream beds may be honeycombed with potholes (Fig. 238).

FIG. 238. A stream bed of limestone honeycombed with potholes. Near Boonville, New York.

CHAPTER IX

GLACIERS AND THEIR WORK

Nature and Significance of Glaciers

When a body of ice, which has been formed from compacted snow, begins to spread or flow from its place of accumulation it is called a *glacier*. In short, a glacier may be defined as a mass of flowing ice on land. Glaciers vary in size from a fraction of a square mile to many hundreds of thousands of square miles. They are found on the earth today in mountains at high altitudes or at lower elevations at high latitudes.

Fig. 239. Northwestern Glacier, Alaska, entering tidewater. (Photo by U. S. Grant for U. S. Geological Survey.)

Much of the land of the earth is, during at least part of the year, covered by snow or ice, and considerable areas are perpetually covered. Moisture, locked up in the form of snow and ice, would tend to accumulate indefinitely in regions of perpetual snow if it were not for the important part played by glaciers in returning much of this moisture to lower and warmer levels.

Glaciers, like rivers, perform their principal geological work by erosion of the land, and by transportation and deposition of rock débris.

Although such work accomplished by glaciers is, on the whole, much less than that of streams, it is, nevertheless, of great importance. Streams have been constantly at work upon most of the lands for tens of millions of years, while glaciers have been much more restricted both in time and place. Water, wind, and ice are the three greatest agents which operate to modify the lands of the earth by the processes of erosion and deposition (gradation).

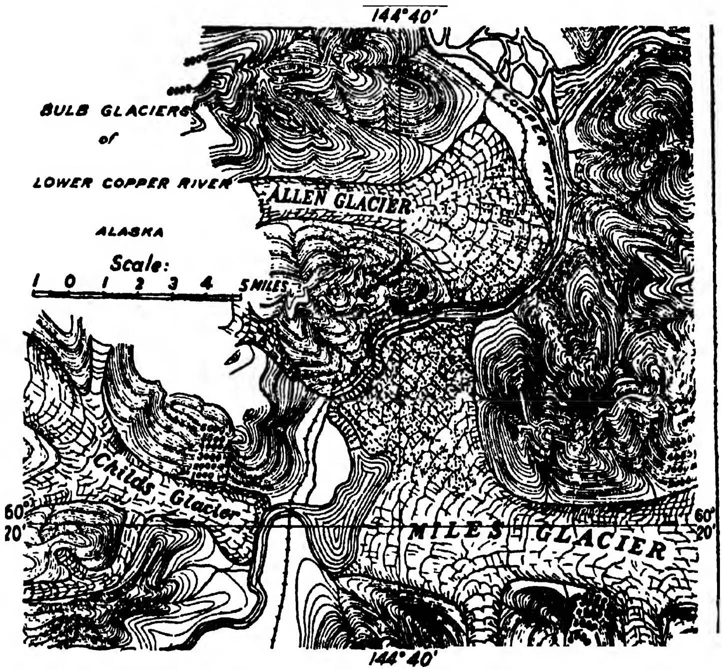

Fig. 240. Map showing how the terminal portions of several large Alaskan glaciers spread out on entering Copper River valley. The full length of Miles Glacier is about 50 miles. (After Witherspoon, U. S. Geological Survey, 1906.)

Types of Glaciers

According to their form, size, and position, we may recognize the following common types of glaciers: (1) *Valley glaciers,* (2) *piedmont glaciers,* (3) *ice sheets.*

Valley Glaciers. These are often called *alpine glaciers* because of their typical development in the Alps where they were first carefully

studied. They are streams of ice flowing through valleys and fed from catchment basins of snow located in regions of perpetual snow. They may have tributaries but, as compared to rivers, these are relatively few in number. Of all the types of glaciers, valley glaciers are the most abundant. They range in length up to about nine miles in the Alps, and to about 75 miles in southern Alaska (Fig. 250). Valley glaciers very commonly attain thicknesses of a few hundred feet to a thousand feet or more, and widths of one-fourth of a mile to several miles. Most valley glaciers are confined to mountain valleys.

FIG. 241. The lower end of Denver Glacier near Skagway, Alaska.

Special Types. A special type of mountain glacier is called a *hanging* or *cliff* glacier. It is a poorly formed, usually small, glacier which occupies a depression or steep cleft high up on mountainsides and does not descend into a valley. It sometimes moves to the edge of a cliff or a very steep slope and breaks off whereupon the fragments which fall to the base of the slope may freeze together again and form a *reconstructed glacier*. The Lefroy glacier near Lake Louise, British Columbia, is a good example (Fig. 243). There are many fine examples of hanging glaciers in the Rocky Mountains of southern Canada and northern United States, and in the Cascade Mountains of Washington and Oregon (Fig. 242).

Hanging glaciers show all stages of transition to true valley glaciers. Such intermediate types are wonderfully displayed on the great volcanic cone of Mt. Rainier, Washington, whose very steep sides support a system of nearly 50 square miles of glaciers (Fig. 244).

FIG. 242. Hanging glaciers near Lake Chelan in the Cascade Mountains of Washington. (Photo by U. S. Reclamation Service.)

Piedmont Glaciers. A piedmont glacier is formed by the coalescence of the spreading ends of valley glaciers where they flow down mountains and out upon relatively level country. It is, in effect, somewhat like a lake of ice at the foot of a mountain. The Malaspina Glacier, covering 1500 square miles at the foot of the great Mt. St. Elias in southern

Alaska, is a fine large example. It has a nearly level surface, and it moves very slowly. Its border portions are almost completely concealed under rock débris and even forest growths. Muir Glacier in Alaska is intermediate in general character between a valley glacier and a piedmont glacier. It covers hundreds of square miles (Fig. 247).

Ice Sheets. These are broad mantles or coverings of glacial ice which are not confined to valleys, but which, instead, bury and spread across topographic forms of various sorts. There are two types, based principally upon size, namely, (a) ice caps and (b) continental ice sheets.

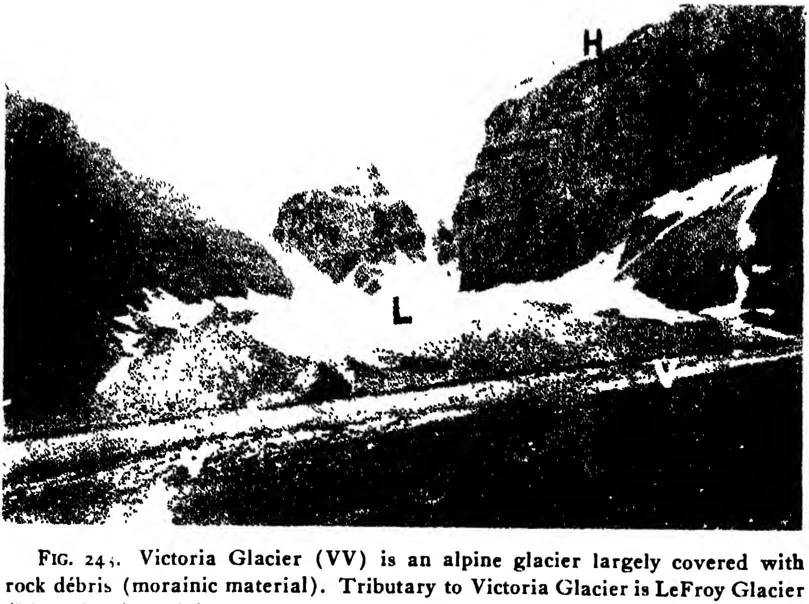

FIG. 245. Victoria Glacier (VV) is an alpine glacier largely covered with rock débris (morainic material). Tributary to Victoria Glacier is LeFroy Glacier (L) which is mainly a reconstructed or débris glacier formed by freezing together of small and large fragments of ice which break from the hanging glacier (H) and tumble down the cliff. Near Lake Louise, British Columbia.

Ice Caps. In certain high-latitude regions such as Scandinavia, Iceland, and Spitzbergen, glacial ice may accumulate on relatively level plains or plateaus as small ice sheets which slowly spread or flow radially from their centers. These are called ice caps. They seldom cover more than a few hundred square miles. If properly situated, they may send small alpine glaciers down radiating valleys.

Continental Glaciers. These are *ice sheets* of great extent, usually covering many thousands of square miles. They are, in principle, much like ice caps, only they are larger. A vast ice sheet now covers fully 700,000 square miles of Greenland, and its motion is outward in all directions toward the sea. It sends off many tongues of ice into the tide

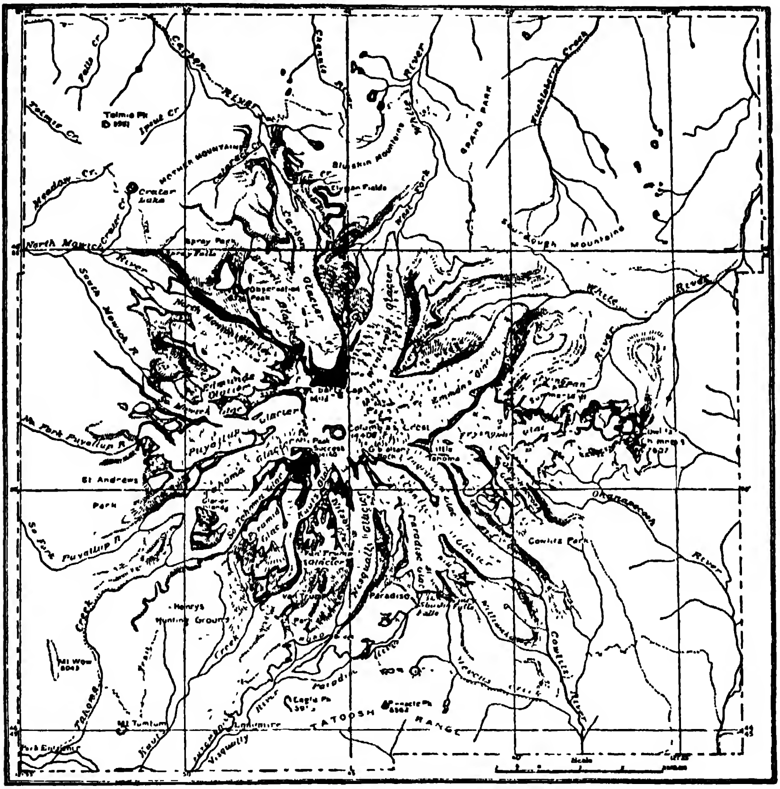

FIG. 244. Map showing the great system of 50 square miles of radiating glaciers on Mt. Rainier, Washington. (After U. S. National Park Service.)

water. A still greater ice sheet covers much of the south polar region to an extent of probably at least several million square miles. The Greenland and Antarctic ice sheets are the only ones at present large enough to be classed as continental glaciers. In times past, however, still greater expanses of glacial ice are known to have occupied certain portions of the earth, as mentioned beyond in this chapter.

Fig. 245. Snow- and ice-clad Mt. McKinley (altitude, 20,300 feet) in Alaska. The main valley is occupied by a large snow-covered glacier. (Photo by R. Muncy, courtesy of J. Danyo.)

Existing Glaciers

Millions of square miles of the earth are covered with glaciers ranging in size from a fraction of a square mile to millions of square miles. Greatest of all are the vast ice sheets, or continental glaciers, occupying much of Greenland and Antarctica. Ice caps of less extent occur in the Arctic Islands, Spitzbergen, Iceland, and southern Scandinavia. Piedmont glaciers, like ice caps, are not very common, their best representation being probably in southern Alaska.

Fig. 246. Andrews Glacier, at an altitude of 12,000 feet in Rocky Mountain National Park. It is a mere remnant of a once large alpine glacier. (Photo by W. T. Lee, U. S. Geological Survey.)

Of all the types of glaciers, the valley or alpine type is by far most abundant. They are best known in the Alps where there are probably no less than 2000 of them. Most of them are less than one or two miles long; a few are from three to five miles long; and one—the Great Aletsch—is over nine miles long. In Europe, the Pyrenees, Carpathian, Caucasus Mountains, and the mountains of Norway also support numerous valley glaciers, those of Norway and the Caucasus being especially large.

The great Himalayas of southern Asia support a magnificent system of very large, high-altitude, valley glaciers, many of them from 5 to 30 miles long. Africa contains few if any glaciers.

The Andes Mountains of South America support many valley glaciers, some small ones at very high altitudes lying practically at the

equator. There are many large valley glaciers in the southern Andes.

There are no glaciers in the eastern two-thirds of North America, but the western portion of the continent contains many of them. There are tens of thousands of glaciers in southern Alaska, most of them by far being valley glaciers which range in length up to 75 miles and in width up to five or six miles. Dozens of them flow down the mountains into tidewater where they break off to form icebergs. Southern Alaska is a wonderland of lofty mountains, vast fields of perpetual snow, and numerous, great valley glaciers (Figs. 245 and 250). Valley glaciers of fair size and hanging glaciers are common in the southern Canadian Rockies. In the northern Rockies of the United States, from Colorado into Montana, there are scores of small glaciers, mostly of the hanging-glacier type, especially in Glacier National Park. The Cascade Mountains of Washington, Oregon, and northern California, especially the higher peaks such as Mt. Rainier, Glacier Peak, Mt. Hood, Mt. Jefferson, and Mt. Shasta, support numerous glaciers ranging from hanging glaciers to true valley glaciers from a fraction of a mile to five miles in length. Some small hanging glaciers occur in the southern half of the Sierra Nevada Range of California. Certain high peaks of Mexico support small glaciers.

The Great Ice Age

The Fact of the Ice Age. The Quaternary is the latest great geologic period of earth history. The first epoch of this period, known as the Pleistocene, was ushered in by the spreading over much of northern North America and Europe of vast ice sheets which must take rank as one of the most interesting and remarkable occurrences of geological time. During several other periods of geological time, glacial ice was more or less extensively developed, particularly during late Paleozoic time, but the term "Ice Age" refers to that of the Pleistocene epoch. Existing glaciers are but remnants of the once much greater glaciers of the Ice Age. On first thought the former existence of such vast ice sheets seems unbelievable, but the Ice Age occurred so short a time ago that the records of the event are perfectly clear and conclusive. The Ice Age is estimated to have begun from half a million to a million or more years ago, and to have ended in the northern United States from twelve to fifteen thousand years ago.

Some of the proofs of the former presence of the great ice sheet are as follows: (1) polished and striated rock surfaces (Fig. 260) which

are precisely like those produced by existing glaciers, and which could not possibly have been produced by any other agency; (2) glacial boulders which are often somewhat rounded and scratched, and which have often been transported many miles from their parent rock ledges (Figs. 277 and 278); (3) true glacial moraines, especially terminal moraines, like the one which extends the full length of Long Island, and

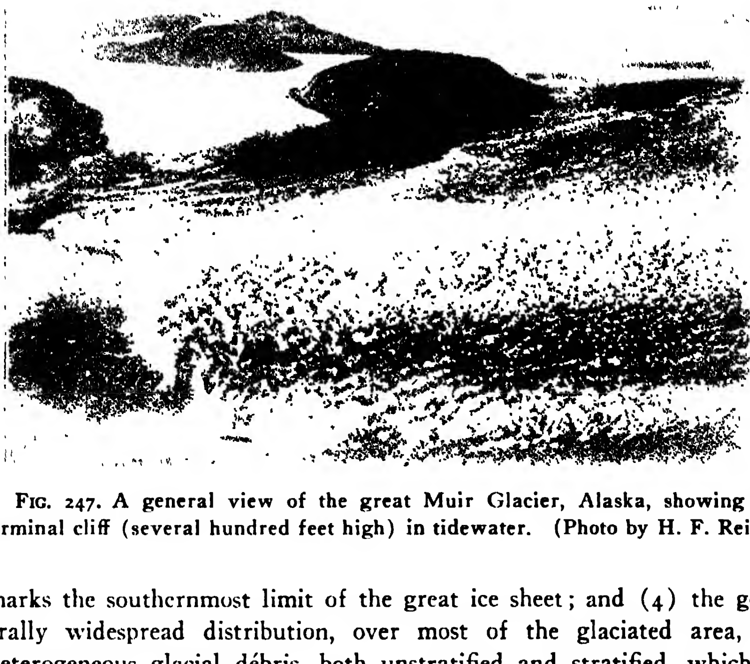

Fig. 247. A general view of the great Muir Glacier, Alaska, showing its terminal cliff (several hundred feet high) in tidewater. (Photo by H. F. Reid.)

marks the southernmost limit of the great ice sheet; and (4) the generally widespread distribution, over most of the glaciated area, of heterogeneous glacial débris, both unstratified and stratified, which is clearly transported material, and which typically rests upon the bedrock by sharp contact (Fig. 270).

Extent, Movement, and Depth of the Ice. An area of about 4,000,000 square miles of northern North America was covered by ice at the time of maximum glaciation. Map Figure 249 shows not only the extent of the ice, but also the three great centers or districts where the glacial ice, compacted from snow, accumulated most abundantly. From each of these three centers—Labradorean, Keewatin, and Cordilleran—the ice

slowly spread in all directions until the three great ice sheets coalesced everywhere except in one relatively small district. This nonglaciated area covers about 10,000 square miles, and lies mostly in southwestern Wisconsin (Fig. 249). It represents a district where the Labradorean and Keewatin Glaciers did not quite join.

The Labradorean and Keewatin Glaciers completely covered the land areas which they invaded. Even the highest mountains of New England and New York were submerged under the ice flood. The

FIG. 248. Asulkan Glacier, British Columbia. The small branch of this glacier, extending across the middle of the picture, is shown in contact with the bedrock over which it is moving.

Cordilleran Glacier did not so completely bury the landscape, many of the highest peaks having projected through the ice. The general depth of the vast glaciers was from one to two miles or more.

A great ice sheet also covered about 2,000,000 square miles of northern Europe during the Ice Age. It radiated from the Scandinavian region, and spread southwestward over nearly the whole of the British Isles; southward into central Germany; and southeastward into central Russia.

The fact that glacial ice flows as though it were a viscous substance is well known from studies of present-day glaciers in the Alps,

Alaska, and Greenland. A common assumption either that the land at the center of accumulation must have been thousands of feet higher, or that the ice there must have been immensely thick in order to permit flowage so far out from the center, is not necessary. For instance, if

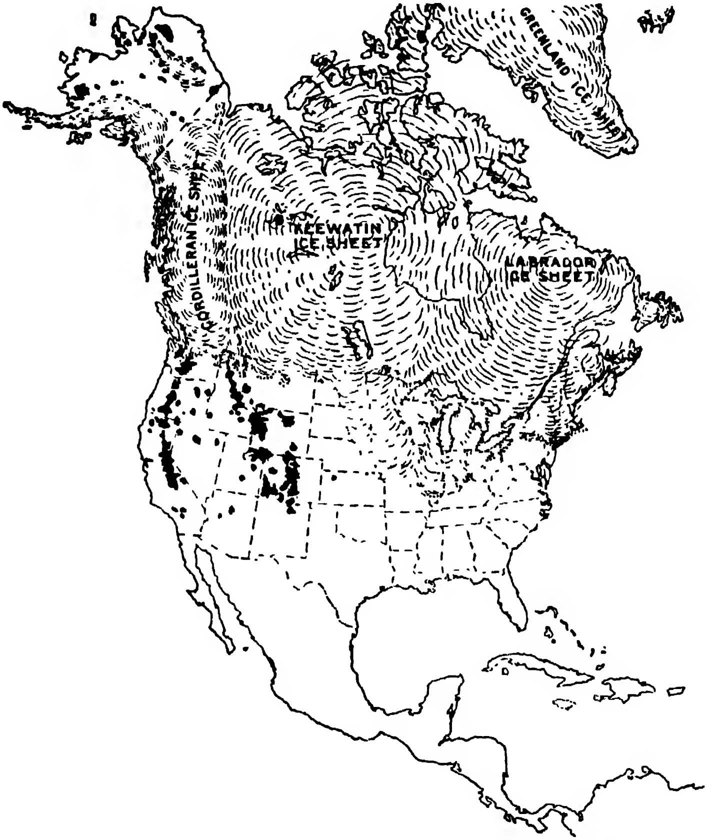

Fig. 249. Map of North America showing the maximum extent of glaciers during the "Ice Age." Areas of local mountain glaciers are shown in black. (Modified after U. S. Geological Survey.)

one proceeds to pour viscous tar slowly in one place upon a perfectly smooth, level surface, the substance will gradually flow out in all directions, and at no time will the tar at the center of accumulation be very much thicker than at other places. The movement of the ice from one of the great centers was much like this, only in the case of the glaciers

FIG. 250. An aerial view of Barnard Glacier (looking northward) in the St. Elias Range, Alaska. The dark bands are superglacial moraines. Note the tributary glaciers emerging from side canyons. The full length of the glacier is about 75 miles. (© Bradford Washburn.)

the accumulation of snow and ice was by no means confined to the immediate centers of accumulation.

Successive Ice Invasions. It has been established that the front of the great continental glacier underwent many more or less local advances and retreats. In the northern Mississippi Valley there is positive proof of at least four important advances and retreats of the ice which gave rise to true interglacial stages. The strongest evidence is the presence

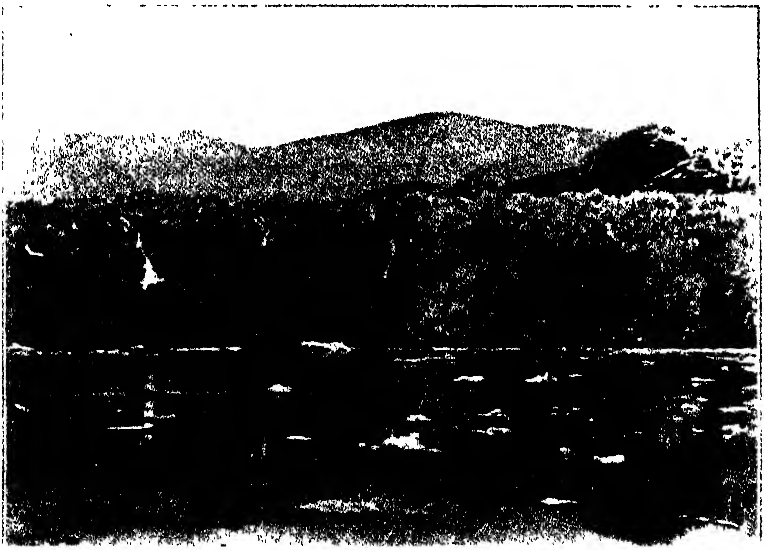

FIG. 251. Detail view of part of the tidewater terminus of Taku Glacier in southern Alaska. Cliff of ice is about 200 feet high. (Photo by Rau Art Studios, Philadelphia.)

of successive layers of glacial débris, a given layer often having been oxidized, eroded, and covered with vegetation before the next (overlying) layer was deposited. In drilling wells through the glacial deposits of Iowa, for example, two distinct deposits or layers of vegetation are often encountered at depths of from 100 to 200 feet. Near Toronto, Canada, plants which actually belong much farther south in a warmer climate have been found between two layers of glacial débris. Thus we know that some, at least, of the ice retreats produced interglacial stages with warmer climate, and that they were sufficient to reduce greatly the size of the continental ice sheet or possibly to cause its entire disappearance.

Origin of Glaciers

Perpetual Snow Fields. Glacial ice is derived from snow. Two conditions are necessary for the formation of glaciers—low temperature and sufficient snowfall. These conditions obtain in perpetual *snow fields,* that is, areas over which the snow persists season after season, and year after year (Fig. 245). In such snow fields there is a tendency for snow to accumulate faster than it can be removed by melting or evaporation, and the excess snow is removed by being transformed into glacial ice as explained below. Some snow fields are too small to produce enough ice for glacier motion.

The line above which snow is always present is called the *snow line.* It is, in other words, the lower edge of a snow field. Snow fields occur in all the regions already mentioned as containing glaciers. They are not uncommon, usually at relatively high altitudes, on all the great land divisions of the earth except Australia, which has none, and Africa, whose few small snow fields are confined to a group of high mountains in the east-central part of the continent.

In the Antarctic and in parts of the Arctic regions the snow line is at or near sea level, while in the equatorial region it is from 15,000 to 18,000 feet above sea level. In certain other parts of the world the altitudes of the snow line (in feet) are approximately as follows: Alps, 9000; Pyrenees, 6500; southern Norway, 5000; Himalayas, 15,000 to 17,000; Bolivian Andes, 15,000 to 18,000; southern Chile, 2000; Mexico, 15,000; Sierra Nevada, 11,000 to 13,000; Cascade Mountains, 8000 to 11,000; Colorado, 12,500; Yellowstone Park, 10,500; Glacier Park, Montana, 9000; southern Alaska, 5000; and southern Greenland, 2000.

Change of Snow into Ice. Every perpetual snow field is also, in part at least, a field of ice. As the snow of such a field accumulates it gradually undergoes a change, especially in its lower portions, first into granulated snow, called *névé,* and then into solid ice. In the late winter and early spring, snow banks in the northern United States often exhibit such a granular appearance. In a snow field the névé grades downward into porous ice and finally into solid ice.

The transformation of snow, through névé, to ice is effected mainly by the weight of overlying snow, which squeezes together and compacts the snow crystals, and by rain or melting snow working down into the snow, there to freeze and fill spaces between the snow crystals.

When the ice beneath a snow field becomes deep enough (usually at least several hundred feet), the spreading action or flowage develops, and a glacier is formed. Repeated falls of snow over the gathering ground of the glacier keep up the supply of glacial ice.

Movement of Glaciers

Rate of Movement. The average rate of movement of glaciers is far less than that of rivers. Many observations have shown that the average rate of movement of the glaciers of the world is not more than a few feet per day. Most of the valley glaciers of the Alps move from one to three feet per day, and this is about an average rate for glaciers of this type. A most exceptional case is a certain glacier, extending as a tongue of the great Greenland ice sheet, whose rate has been found to be 60 to 70 feet per day. Some of the very large glaciers of Alaska move at rates of from 4 to 40 feet per day. A glacier advances across country only when its rate of movement is greater than its rate of melting.

Laws of Glacier Motion. The nature of glacier motion is by no means simple. It involves *differential motion* in a rather complex sense of that term. Brief mention of most of the so-called "laws of glacier motion" will serve to make clear the complicated nature of the movement. These laws, which apply most typically to valley glaciers, are as follows:

1. A glacier, to a greater or less extent, actually glides or slides over the earth's surface. This is abundantly proved by the eroded, and often polished and striated, rock surfaces left by glaciers.

2. The top portion of a glacier moves faster than the bottom, because of friction of the glacier on its bed. This has been proved by observing the change in position of a vertical line of pegs driven into the steep side of a valley glacier.

3. The middle portion moves faster than the sides because of friction of the glacier against its containing banks. This is easily proved by observing the changing position of a row of marked objects placed across a valley glacier.

4. The velocity increases with steepness of slope of the bed. This has been proved particularly for certain glaciers in the Alps. It must be so because gravity is the ultimate force which causes the motion.

5. The velocity increases with the thickness of ice. This again is

due to the fact that the force of gravity is more effective in causing movement if a body of glacial ice on a slope is relatively thick.

6. The velocity increases with temperature. In warm weather a glacier moves faster than in cooler weather, that is, moves faster when it is melting and contains more water.

7. Velocity increases with straightness of course. A glacier flows less rapidly through a crooked valley because the friction is greater as the ice current rounds the curves.

8. Velocity diminishes with roughness of bed. The motion of the glacier is retarded by being forced over inequalities or obstacles in its bed.

9. Velocity diminishes with amount of load in the basal portion. This is because of increased friction of the glacier on its bed.

10. The line of greatest velocity is more winding than that of the glacial channel. Just as in a river, the tendency also in a winding glacier is for the line of greatest current to swing back and forth from one side to the other.

11. A stream of ice does not conform to minor irregularities of the sides of the channel. A glacier several hundred feet thick may move past the end of a tributary valley without flowing into the latter to seek the general ice-level. In this respect, glacier movement is very different from that of water.

Except for the force of gravity which inaugurates the movement, the cause of glacier motion is not yet definitely known. Several theories have been advanced, but it would carry us into too great detail to discuss them in this book.

Lower Limits of Glaciers

We have already learned that glaciers almost invariably originate in regions of perpetual snow. A rare exception to this rule might be the formation of a reconstructed glacier below the snow line. Under favorable topographic conditions, most glaciers of considerable size flow down to greater or less distances below the snow line. This is particularly true of valley glaciers. Many small hanging glaciers move little if any below the snow line. Piedmont glaciers generally form well below the level of the snow field. Ice caps often send tongues of ice below the edge of the snow field. Continental glaciers usually lie very largely within snow fields, though around their borders the ice may extend beyond the snow line.

FIG. 252. The Illecillewaet Glacier, British Columbia, as it appeared in 1912. (Photo by H. Ries.)

FIG. 253. The Illecillewaet Glacier, British Columbia, as it appeared in 1921. Compare with Figure 252, and note the amount of retreat of the ice.

Valley glaciers not uncommonly move some miles beyond and several thousand feet below the line of perpetual snow. A comparison of altitudes of some examples of lower limits of glaciers with the altitudes of the snow line in the same regions as above listed will be instructive in this connection. In the southern Sierra Nevada Range of California the lower limit of glaciers is about 12,500 feet. On Mt. Shasta in northern California it is about 9000 feet. The lower limit in the Cascade Mountains of Washington is about 4500 feet, while at the same latitude in the Rocky Mountains of Montana it is about 6500 feet, the difference being due mainly to the greater snowfall in the former region. In southern Alaska a number of the great glaciers move down to tide water, there to break up in the form of icebergs. Glaciers of southern Greenland reach the sea. In the Alps the lower limit is about 4000 feet. A remarkable case is in New Zealand where large glaciers on South Island flow down into sub-tropical forests of tree ferns.

The position of the lower end of a glacier depends upon the relation between rate of movement and rate of melting and evaporation of the ice. When rate of movement predominates over rate of evaporation and melting, the end of a glacier advances, and vice versa. A rather delicate balance will cause the end of a glacier to remain stationary for a time. A series of seasons of heavy snowfall over the gathering-ground of a glacier will in time cause advance, whereas a series of seasons of light snowfall will cause retreat. "There are reasons for believing it probable that there are cycles of advance and recession, due, perhaps, to climatic variation; and careful records are now being kept in the hope of discovering the cause for variations in the position of ice fronts" (Tarr and Martin).

Most of the glaciers of Europe and North America are now retreating. Thus the Rhone Glacier in the Alps has retreated a considerable fraction of a mile in the last 40 years. The Illecillewaet Glacier in the Selkirk Mountains has retreated hundreds of feet during the last 30 years (Figs. 252 and 253). Nisqually Glacier on Mt. Rainier, Washington, retreated 1864 feet from 1919 to 1944. The tidewater front of the great Muir Glacier of Alaska has retreated several miles in the last 35 years.

Features of Glaciers

Crevasses. The surface of a glacier is usually very rough, irregular, and broken, often making travel over it difficult or even dangerous. The roughness is due in part to irregular melting of the ice; to streams

of water which melt and erode channels in the ice; and to irregular accumulations of rock débris, called moraines, described beyond. The major irregularity and roughness of surface are, however, due to the presence of numerous small and large cracks and fissures which will now be described and explained. These cracks and fissures are of three general types. They vary in width up to 20 feet or more and in depth to hundreds of feet.

Fig. 254. A great transverse crevasse in South Sister Glacier, Cascade Mountains, Oregon. (Courtesy of U. S. Forest Service.)

Much like molasses candy, ice tends to crack when subjected to a relatively sudden force, particularly a force of tension. Thus where there is a rapid increase in slope across the bed of the glacier, the ice may not be able to mold itself over the salient without rupture, and *transverse crevasses* develop across the glacier (Fig. 254). This is because tension in the upper portion of the glacier is greater than in the lower portion over the salient in the bed (Fig. 255). A rapid change of slope of only a few degrees is usually sufficient to cause transverse

crevasses. Because of the forward motion of the ice, old crevasses often close up, and new ones develop over the salient.

As a result of the greater velocity of the central portion of a valley glacier, stresses and strains set up in the marginal portions often cause *marginal crevasses* to develop. Such cracks usually extend obliquely upstream from each margin of the ice well into the glacier at angles of approximately 45° (Fig. 255).

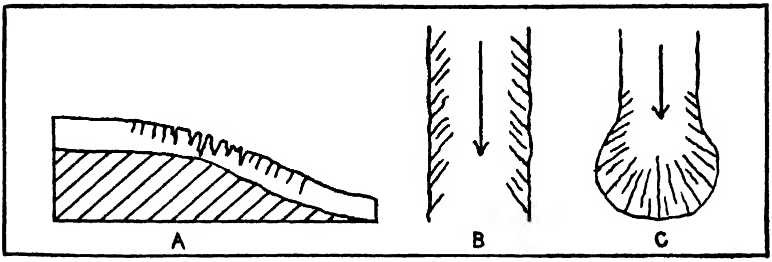

FIG. 255. Sketches showing origin of crevasses in glaciers. A, structure section showing transverse fissures; B and C, ground plans showing marginal and longitudinal fissures.

Where a glacier spreads laterally in a broader portion of a valley, or where it terminates and spreads out on a nearly flat surface, *longitudinal crevasses,* that is, cracks roughly parallel to the direction of ice-flow, usually develop. In such cases the ice by fracturing yields to the force of tension which is caused by relatively rapid spreading (Fig. 255). Crevasses do not extend far down in thick glaciers because the ice, under much pressure at depth, is more subject to flowage.

A type of crevasse not really within the body of the glacier should be mentioned. This is the *bergschrund* which develops at the head of the glacier where the glacier motion begins. This fissure (or series of them) forms where the thick body of ice, névé, and more compacted snow of the snow field draw away from the thinner, less compacted snow of the upper margin of the snow field (Fig. 256). The bergschrund will be referred to again in the discussion of glacier erosion.

Stratification. The ice of a glacier is often crudely stratified. This is because the ice of the snow field is built up of successive falls of snow, each of which has certain more or less characteristic features of texture, compactness, depth, etc. When changed into ice, some of the *layers* are more porous and white than others which are very compact

and blue. Also, during considerable intervals between falls of snow, especially between the seasons, more or less dirt often accumulates on the surface of the snow field. Such dirt bands greatly accentuate the stratified appearance of the ice which is, as a rule, best developed toward the upper end of the glacier. It is, however, often more or less noticeable even at the very end of a glacier (Fig. 258) in spite of the complicated motions to which the glacial ice has been subjected.

Fig. 256. A bergschrund at the head of a glacier. Swiss Peak, British Columbia. (Photo by L. G. Westgate.)

Veinlike structures are often developed locally where porous ice is subjected to an extra degree of compression during the glacier movement, causing it to become solid (and blue) by squeezing out the air.

A crude stratiform or layered structure also develops, especially in the deeper parts of a glacier, by the sliding or shearing of one part of the ice over another. Such sheared surfaces may at times look something like stratification surfaces.

Moraines of the Ice. Most glaciers carry or drag along more or less rock débris ranging in size from very finely divided material to great boulders. Such débris is transported by a glacier either on its surface or within it or in or under its bottom portion. The term *moraine* applies to all material gathered, transported, and deposited by glaciers. Morainic material is represented partly by rock fragments, which are rolled or washed down upon the glacier, and partly by rock fragments eroded by the glacier from the bed and sides of its channel. Morainic material carried on top of the glacier may be called *superglacial;* that frozen within it, *englacial;* and that in and just under its bottom portion, *subglacial.* Upon melting of the glacier, it forms moraines on the ground where deposited, (to be discussed later).

The superglacial débris is mostly of two classes—lateral and medial. Where it is arranged along the sides of the glacier it is called a *lateral*

moraine. It consists mainly or wholly of material which has rolled or washed down upon the margins of the glacier from its bordering rock walls or sides. It is usually most conspicuous toward the end of the glacier where it forms ridges of earth from a few feet to a hundred feet or more high. A *medial moraine* is a belt of rock débris on the surface of the glacier, well away from its margin. It may or may not be in the middle of the glacier. It nearly always results when two glaciers flow together, so that two adjacent lateral moraines (one from each glacier) unite to form a medial moraine (Fig. 257). A trunk glacier, formed by the union of several tributaries, may show several medial moraines.

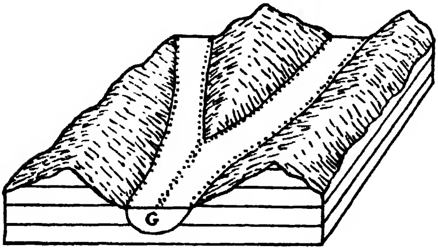

FIG. 257. Block diagram showing moraines (dotted portions) on a glacier (G). Two of the lateral moraines of the tributary glaciers join to form a medial moraine.

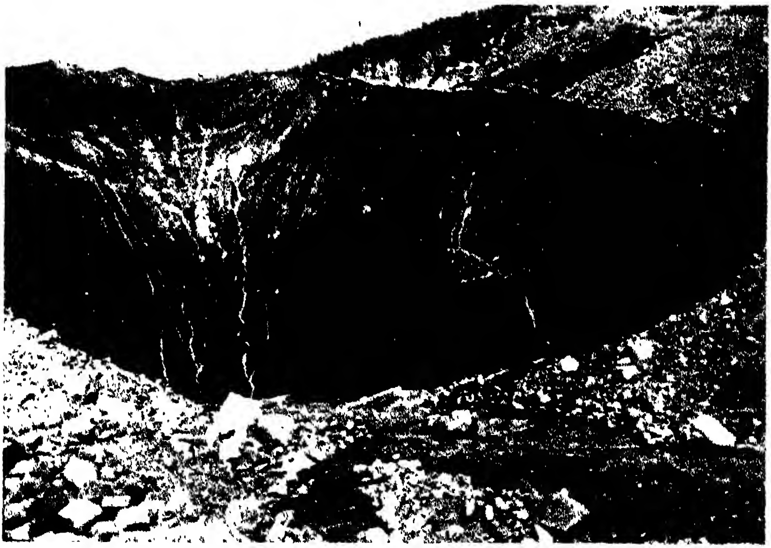

FIG. 258. Terminus of the Nisqually Glacier, Mt. Rainier, in August, 1931. The ice is dark because of the large amount of englacial rock fragments contained in it. Note the stream emerging at the bottom and the layered or banded structure at the upper left.

FIG. 259. An aerial view looking southward over the Convict Lake area on the eastern front of the Sierra Nevada Range, California. The lake (altitude 7583 feet) lies in a deep basin once occupied by a thick glacier which flowed out upon the valley floor for about one and one-half miles as indicated by the moraines. Terminal moraine, T; lateral moraines, L, L; and recessional moraines, R_1, R_2, R_3. The outlet creek has cut through the moraines on the left side. (By Fairchild Aerial Surveys, Inc.)

Certain interesting topographic features of the surfaces of glaciers result from the influence of the superglacial moraines. Not only do such moraines often form ridges by their accumulation, but also (when thick enough) they protect the ice immediately underneath them against melting and evaporation, by which processes the general surface of the glacier is very appreciably lowered. In this manner the morainic ridges are accentuated in height. For the same reason, large blocks of rock may be left perched temporarily upon ice pedestals or columns. A very thin surface layer of rock débris absorbs enough heat to cause the ice just underneath it to melt faster than otherwise, and so depressions of various shapes and sizes result.

Englacial material results partly from rock débris which accumulates on the surface in the catchment basin and is buried under new falls of snow which change to ice, and partly from débris which falls into crevasses in the glacier farther down its course. Englacial material may travel miles through the body of a glacier and then emerge at or near its terminus. Marked objects thrown into the sources (catchment basins) of glaciers many years ago have been found to emerge at or near the lower ends of the glaciers. When, through melting and evaporation of the top ice, some of the englacial material appears at the surface of the glacier, it becomes *superglacial material.*

Subglacial material, also called the ground moraine, is lodged within, or dragged along just under, the bottom of a glacier. It consists of superglacial and englacial materials which make their way to the bottom, together with materials picked up by glacier erosion. The greatest portion of all morainic material is carried in the bottom portion of a glacier.

All rock débris—superglacial, englacial, and subglacial—carried along by a glacier ultimately tends to reach its terminus where it accumulates to form the *terminal moraine.* Such a moraine becomes most conspicuous when the terminus of the glacier remains practically stationary for some time (Fig. 259).

Drainage of Glaciers

In mild weather, a glacier nearly always has streams of water upon it. Most of this water results from melting of the ice, but some of it may flow down the valley sides and thence upon the glacier. Most of these streams are very temporary, and they usually do not flow far before pouring into crevasses or over the sides or end of the glacier. Some of the water follows englacial channels for a time, and some of

this englacial water may issue from the sides of the glacier above its bottom in the form of springs. The general tendency is, however, for the water to accumulate in the form of a stream at the bottom of the glacier and to issue at or near the terminus of the latter, often from a tunnel (Fig. 258). The water of such a subglacial stream is characteristically turbid and whitish because it is charged with very finely ground particles of fresh, unweathered rock. Ordinary streams in flood are usually brownish or yellowish because charged with weathered material rich in oxide of iron.

Glacial Erosion

How Glaciers Erode. Glacial ice can accomplish more or less erosion of loose and soft rock materials. But, like running water, ice has considerable power to erode relatively hard rock only when it is properly supplied with tools.

Fig. 260. A glaciated ledge of sandstone high up in the Rocky Mountains. Glacier National Park, Montana.

An important process of glacial erosion is *abrasion*, that is, the rubbing and grinding action of rock fragments either frozen into the bottom and sides of the glacier or situated just underneath it. Much of the work of erosion is, then, accomplished not by the ice itself but rather

by the rasping, grinding, and rubbing action of the rock fragments carried along by the glacier. Rock surfaces which have been subjected to glacial abrasion are characteristically smoothed and usually more or less scratched, striated, or grooved (Fig. 260). Such scratches and grooves are known as *glacial striae*. A glaciated rock surface of this kind constitutes one of the best proofs of the former presence of a glacier in a region, and the striae indicate the direction of the glacier movement.

FIG. 261. A ridge of granite strongly scoured from right to left by a glacier. Glacial boulders in the foreground. Tuolumne Meadows, California.

Another important process of glacial erosion is *plucking* or *quarrying*. This consists in separating from the bedrock, and pushing along, blocks of rock already more or less loosened by joint cracks. Highly jointed rocks are, therefore, most susceptible to glacial plucking. Such joint blocks, as well as any other boulders and pebbles, which are rubbed against either the bedrock or each other by the movement of the glacier, often become faceted and striated.

Significance of Glacial Erosion. Considering the present and past condition of the earth, the total work of ice erosion as compared to that of running water is slight, because glacial erosion is, and has been, much more restricted in its action both in space and time. During the last 50 years various opinions have been expressed in regard to the efficacy of glacial erosion. Some geologists have ascribed great erosive power to glaciers, whereas others have considered them to be weak erosive

FIG. 262. A U-shaped glaciated canyon in the Rocky Mountains. Swiftcurrent Valley, Glacier National Park.

FIG. 263. Looking eastward through the deep, steep-sided Yosemite Valley, California. El Capitan on the left and the Cathedral Rocks and Bridalveil Falls on the right. (Photo by Putnam Studios, Los Angeles, California.)

agents. The present consensus of opinion is that, under reasonably favorable conditions, glaciers accomplish a truly important work of erosion.

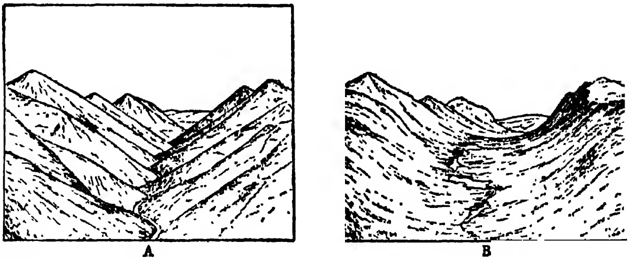

FIG. 264. Diagrams showing a stream-cut valley (A) as it appears after glaciation (B). (After U. S. Geological Survey.)

FIG. 265. An aerial view of part of the western slope of the Sierra Nevada Range, including Yosemite Valley. A deep V-shaped canyon was here changed into a U-shaped glacial trough by the action of a great stream of ice (Fairchild Aerial Surveys.)

When a thick glacier, shod with numerous fragments of hard rock, moves over relatively soft or highly jointed rock, conditions for ice erosion are exceptionally favorable because the grinding tools are hard and abundant, the work to be done is easy, and the pressure is great.

We have no evidence that large valleys have been developed or that the general landscape has been profoundly altered by glacial erosion, but we do have positive evidence that many valleys and landscapes have been notably modified by glacial erosion.

Fig. 266. A typical cirque about 3000 feet deep west of Long's Peak, Colorado. Massive granite has here been excavated.

The efficacy of ice erosion over a large territory is particularly well shown by the effects of the passage of the vast Labradorean Glacier over southeastern Canada, northern New England, and northern New York. Common and characteristic features of this territory are the almost total absence of residual soils, which must have been very abundant before the advent of the ice, and the large number of rounded, glaciated, bare surfaces of very hard and fresh rocks. The evidence is clear and conclusive that the glaciers, shod with many fragments of hard rock, eroded not only practically all the soil and rotten rock from these surfaces but also at least some of the fresh rock. This is true even at altitudes of one to several thousand feet in many places.

Yosemite Valley. A remarkable example of the power of a valley glacier to erode very hard rock is the famous Yosemite Valley of California. This valley is seven miles long, several thousand feet deep, broad-bottomed, and bounded by precipitous walls (Fig. 263). The rock is wholly hard granite or closely related plutonic rock.

Fig. 267. Iceberg Lake and a remnant of a once large valley glacier at the bottom of a great cirque with walls 3000 feet high. The rocks are nearly horizontal strata. Glacier National Park. (Photo by M. R. Campbell for U. S. Geological Survey.)

Just before the Ice Age, the site of Yosemite Valley was a V-shaped canyon about 3000 feet deep cut by the Merced River. During the Ice Age two glaciers, flowing from the high Sierra Nevada, joined to form an extra large glacier which flowed through the Yosemite part of the Merced Canyon notably modifying it, by glacial erosion, into the Yosemite Valley (Fig. 265). Some of the principal factors aiding the ice erosion were: the great thickness (several thousand feet) of the glacier causing high pressure at the bottom and lower sides; the large number of fragments of hard rocks with which the glacier was shod, thus facilitating the work of abrasion; and the existence of many prominent, vertical joints in the granite, thus allowing the plucking action of the glacier to be very effective in removing extensive joint slabs one after

another from the valley sides. When the ice disappeared, the great, nearly vertical joint faces were left much as they appear today. The wonderful, high waterfalls, including the Upper Yosemite Fall (1430 feet), are due to the fact that post-Glacial streams in various places plunge over high, joint-face walls. Glacial débris, in the form of a terminal moraine, so blockaded the drainage at the western end of the valley that a lake, several hundred feet deep, covered the bottom of the valley after the ice melted away. The lake filled with sediment, and this explains the wide, flat alluviated floor of the valley. Once or twice after the big glacier disappeared from the valley, much smaller glaciers occupied it.

Characteristics of Glaciated Valleys. A mountain valley through which a thick glacier has flowed in relatively recent times shows certain unmistakable evidences of the former presence of the ice.

(1) A valley which has been vigorously glaciated has a broad bottom and very steep to vertical sides. In other words, it has a U-shaped cross section or profile instead of the characteristic V-shaped cross section of a valley vigorously eroded by a stream. This is because a glacier not only erodes the bottom of its valley, but also because it actively cuts back its sides, especially toward their bottoms where the ice pressure is greatest. Thus the valley is deepened, its sides are steepened, and its bottom is much broadened (Fig. 264). Swiftcurrent Canyon in Glacier National Park, with its steep walls and broad floor, is a fine example (Fig. 262). There are many other examples in the Sierra Nevada, Cascade, and Rocky Mountains, and in Alaska.

(2) Many of the glacial valleys (or canyons) in the mountains just mentioned are much straighter and more open for long distances than stream-eroded valleys (or canyons) would be. This is because a glacier has a much stronger tendency to take a straighter course than has a river, and so the lower ends of the ridges (spurs) which project down into the valley alternately from opposite sides are truncated by glacial erosion (Fig. 264).

(3) We have already learned that stream-cut tributary valleys very typically join their main-stream valley *at grade,* that is, at practically the same level. In a glacial valley, however, the tributary valleys show typically a *discordance* of position, that is, they join the main valley much above its bottom and are, therefore, called *hanging valleys.* This is because the lower ends of the tributary valleys are cut back during the process of valley widening by the action of the glacier in the main valley.

Even if glaciers occupy the tributary valleys, they are usually too small to cut down their beds as fast as the main glacier. Streams in such tributary valleys usually enter the main-valley streams by waterfalls or cascades. Hanging valleys with waterfalls are grandly exhibited in the Yosemite Valley. Many other examples occur in the Sierra Nevada, Cascade, and Rocky Mountains (Fig. 268), and in southern Alaska, Norway, and the Alps.

FIG. 268. A hanging cirque opposite Going-to-the-Sun Chalet, Glacier National Park.

(4) The heads of glacial valleys are very commonly characterized by big, steep-sided, amphitheater-like basins known as *cirques*. As we have already mentioned, the main body of snow, névé, and ice of the snow field at the head of a valley glacier tends to pull away from the snow and névé of the upper slopes, leaving a deep crevasse called the *bergschrund* in which the bedrock is more or less exposed. During the warm days of summer, water fills the joint cracks and crevices in the rocks down in the bergschrund, and during the much colder nights this water freezes, forcing apart and loosening some of the joint blocks. Such rock fragments accumulate in the bottom of the bergschrund where, in the later colder season, they are enveloped in ice and in névé

FIG. 269. A field strewn with glacial boulders or erratics. Near Northampton, Massachusetts.

formed from new snowfalls. The rock fragments are thus frozen into the head of the glacier and carried along by it. This quarrying or excavating operation, which is repeated season after season, is most effective toward the bottom of the bergschrund, and so the sides of the valley head are cut back and greatly steepened, forming a cirque.

Cirques, now free from ice or nearly so, are abundant in the Sierra Nevada, Cascade, and Rocky Mountains (Fig. 267), in southern Alaska, and in the higher mountains of Europe. They are commonly

FIG. 270. Detail view of ground moraine material left by a great glacier of the Ice Age. Adirondack Mountains, New York.

from one-fourth of a mile to a mile or more wide, with steep to precipitous walls from 500 to 3000 feet high (Fig. 266). Occasionally cirques occupy the positions of hanging valleys, excellent examples occurring in Glacier National Park, Montana (Fig. 268). Cirques constitute striking features of the landscape in these and other recently glaciated mountains. They often contain small lakes (Figs. 267 and 365).

Two cirque walls may be cut back toward each other from opposite sides of a mountain mass until only a very sharp divide, known as a *knife-edge* ridge, is left between the cirques. A knife-edge ridge may also develop where glaciers in two parallel valleys erode and steepen the valley sides until only a very sharp divide separates the valleys. If three or more heads of glaciers cut cirques into a mountain mass from several sides at once, a high, pyramid-shaped rock mass, commonly called a *matterhorn,* may result. The type example is the famous Matterhorn of the Alps. Matterhorns are common in Glacier National Park, Montana (Fig. 273).

(5) A less common feature of glacial valleys is that large glaciers entering the sea may erode their valleys hundreds of feet below sea level. This is because the moving ice is able to displace the water until

FIG. 271. Map of the Great Lakes region showing the directions of ice movement and the recessional moraines of the final ice retreat during the Ice Age. (After Taylor and Leverett, U. S. Geological Survey.)

its depth becomes so great that the ice is buoyed up and broken off. A number of large Alaskan glaciers which enter arms of the sea are now at work deepening their valleys hundreds of feet below tide level. Rivers can cut their channels but very little below sea level.

In this connection mention should be made of certain deep-water,

narrow arms of the sea with high steep walls, called *fiords*. They are exhibited on grand scales in Norway and southern Alaska where they are often 10 to 75 miles long and several thousand feet deep; they are also found, on a less grand scale, in Maine. All or nearly all of them

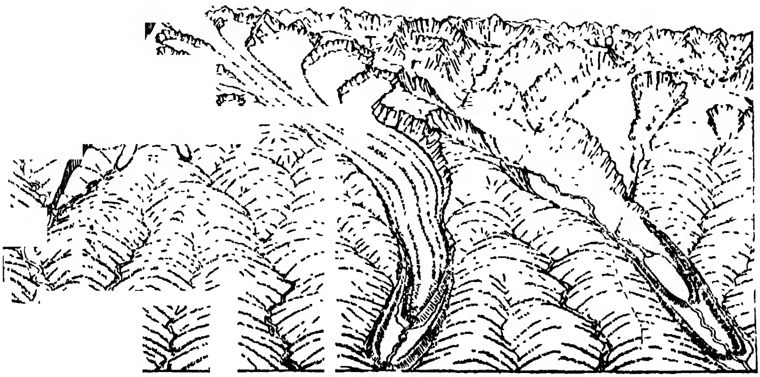

FIG. 272. A three-stage diagram illustrating the glacial sculpture of mountains. On the extreme left, a nonglaciated mountain region; in the middle, similar mountains being modified by glaciers; and on the right, similar mountains after the glaciers disappeared. Note the U-shaped valleys, hanging valleys, cirques, terminal and lateral moraines, and glacial lakes. (After W. M. Davis.)

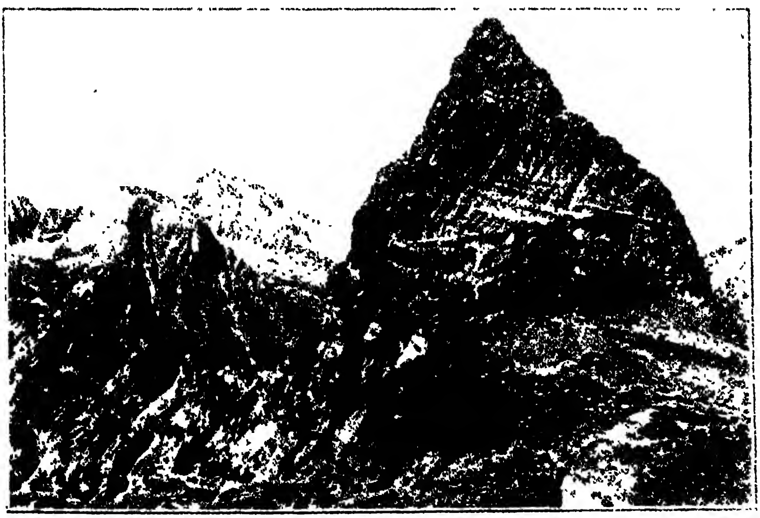

FIG. 273. Matterhorn Peak in Glacier National Park. (Copyright photo by R. E. Marble, Glacier National Park, Montana.)

have resulted from erosion of river valleys by glaciers followed by notable subsidence of the land. They are usually too deep to be accounted for by glacial erosion alone. The maximum depth of water in Norwegian fiords is commonly from 1000 to 4000 feet.

Glacial Deposits

The Drift. We have already learned that glaciers transport large quantities of rock débris either on their surfaces or within them, or dragged along at their bottoms. It is heterogeneous material ranging from the finest clay, through sand and gravel, to boulders weighing many tons. Some of these materials may be deposited during the slow advance of a glacier, but such materials are again very largely eroded and carried along by the advancing ice. Most of the deposits, not again disturbed by the ice, are laid down during the retreat of a glacier, and these are of chief interest to us because they are the ones which are so widespread as a direct result of the recent great Ice Age. Most of the deposits left by the glaciers of the Ice Age are remarkably intact except for relatively little post-Glacial weathering and erosion.

The term applied to all deposited rock material directly or indirectly of glacial origin is *drift.* It is so called because, before the discovery of the fact of the Ice Age, such deposits were regarded as having been carried (or drifted) over the country by floods and icebergs. Much of Canada and most of the northern United States as far south as New York City, Pittsburgh, St. Louis, Pierre, and northern Idaho, are covered by drift from a few inches to several hundred feet thick. It is not shown mainly where bedrock is exposed; where lake and river waters are present; or where there are post-Glacial river and lake deposits. The drift bears unmistakable evidences of its glacial origin. Some of its material has been transported hundreds of miles, as proved by tracing certain types of rocks in the drift to their parent ledges. An important characteristic of the drift is that it rests upon the bedrock by sharp contact, and usually contains at least some material different from the bedrock, showing that it is transported and not residual material.

There are in general two classes of glacial drift, namely, the unstratified *ice-laid deposits,* known as *till,* which are left by the ice unaided by the action of water, and the stratified *fluvio-glacial deposits* which are carried and deposited by waters under, or emerging from, glaciers.

Ice-laid Deposits or Till. *Ground moraine.* This is a sheet of heterogeneous, unstratified rock débris which was deposited underneath a glacier especially during its melting and retreat. When it is mostly very fine material with pebbles or boulders scattered through its mass, it is sometimes called *boulder clay* (Fig. 270). The pebbles and boulders are often characteristically faceted and striated as a result of having been rubbed and ground against the bedrock. Ground morainic material is exceedingly widespread in the great glaciated region of North America.

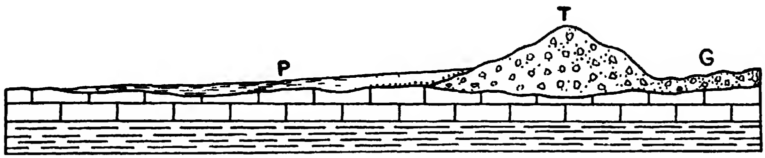

FIG. 274. A structure section showing several kinds of glacial deposits resting upon bedrock. Ground moraine, G; terminal moraine, T; and outwash plain, P.

Terminal Moraines. Whenever the terminus of a glacier remains in a relatively stationary position for a considerable time much rock débris carried by the glacier accumulates around its end, forming a *terminal moraine* (Fig. 259). Such a moraine, especially as left by the great ice sheets of the Ice Age, is a more or less distinct range of low hills, with depressions between the hills, consisting of very heterogeneous, generally unstratified débris, although at times waters emerging from the glacier may have caused some local stratification. Valley glaciers often leave looplike terminal morainic ridges across valleys or at their mouths (Fig. 259). A great terminal moraine is more or less clearly traceable across the United States where it marks the southernmost limit of the vast glaciers of the Ice Age. On Long Island it is wonderfully well shown by the ridge of irregular hills extending the whole length of the island.

If, after a glacier has retreated a considerable distance, its terminus again remains relatively stationary for some time, another terminal deposit will accumulate around it. Such a *recessional moraine* (Fig. 259) develops during every considerable pause in the recession of a glacier. Recessional moraines, forming a great succession of curving ridges, are wonderfully displayed south of Lakes Michigan and Erie (Fig. 271). These mark successive pauses of the waning Labradorean ice sheet of the Ice Age.

Lateral Moraines. When a glacier melts, a *lateral moraine,* formed alongside the living glacier, may be left in the form of a more or less conspicuous ridge (Figs. 259 and 275).

Fig. 275. Part of a lateral moraine recently left by the retreating Nisqually Glacier, Mt. Rainier, Washington.

Drumlins. These are unstratified glacial deposits of unusual interest. They represent, in reality, only a special form of ground morainic material. They are typically low, rounded mounds or hills of till with elliptical bases; long axes parallel to the direction of the glacier movement; and steepest slopes facing the direction from which the ice flowed. They are commonly from 50 to 200 feet high, and one-fourth to one-half of a mile long. One of the grandest displays of drumlins in the world is in the general region between Syracuse and Rochester, New

Fig. 276. Side view of a drumlin. Near Weedsport, New York.

York, where thousands of them rise conspicuously above the level of the Ontario Plain (Fig. 276). Drumlins are abundant in eastern Wisconsin and in a part of Ireland.

The mode of origin of drumlins has not been precisely determined, but it is known that they form near the margins of broad lobes of glacial ice probably either by ice erosion and rounding-off of till or by accumulation of till beneath the ice under favorable conditions, as perhaps in longitudinal crevasses.

FIG. 277. A delicately balanced glacial boulder (so-called "rocking stone") 1500 feet above sea level. Near Westhampton, Massachusetts.

Erratics. These are glacial boulders left strewn irregularly over the country during the melting of the ice. They vary in size from pebbles to masses as big as small houses. Most of them consist of hard rock, because the softer materials are generally ground up soon by the action of the glacier. Some erratics have been moved but short distances from their parent ledges; many have been transported at least a few miles; while some have been carried hundreds of miles. Thus boulders of Adirondack Mountain rocks occur in southern New York, and certain erratics in southern Minnesota came from ledges well up in Canada. Erratics are extremely abundant in New England and New York where much land had to be cleared of them (Fig. 269). They occur high up

on mountains. Erratics weighing from 5 to 20 tons have sometimes been left in such remarkably balanced positions on bedrock that they can be made to swing back and forth slightly by pressure of the hand (Fig. 277). Such boulders are sometimes called "rocking stones." The writer recently observed a large erratic standing on edge at the very summit of a peak 2600 feet above sea level in northern New York. Still another case observed by the writer was a rounded erratic about 14 feet in diameter remarkably balanced on top of another rounded erratic of about the same size (Fig. 278).

FIG. 278. A big glacial boulder remarkably perched upon another boulder. East of Blue Ridge P. O., New York.

Fluvio-glacial Deposits. *Valley trains.* Waters emerging from the ice are usually heavily loaded with rock débris. When such waters flow down a valley which slopes gently downward away from the end of a glacier, the tendency is to deposit some or most of the load on the valley floor, often for miles beyond the ice front, forming a valley train (Fig. 279). Deposits of this kind, somewhat cut away by post-Glacial erosion, are finely exhibited in most of the gently southward sloping valleys of southwestern New York. Valley trains are crudely stratified.

Outwash Plains. When the front of a great glacier pauses for a considerable time upon a rather flat surface, the débris-laden waters emerging from the ice spread in a network of streams and deposit the débris more or less uniformly over the surface, forming an ***outwash plain*** (sometimes called a *frontal apron*) (Fig. 274). A very fine illustration is most of the southern half of Long Island lying just south of the great terminal moraine. Outwash plains are, of course, stratified. They are seldom formed by ordinary valley glaciers.

Depressions from 10 to 100 feet deep, with no outlets and with steep sides, are often formed in outwash plains. These so-called ***kettle holes*** result from the melting of blocks of ice which become separated from the glacier during its retreat and buried in the outwash material. Kettle holes may also develop in glacial lake deposits where icebergs become stranded, buried under sediment, and later melted. Some depressions in moraines and also in groups of kames are kettle holes.

Fig. 279. A broad valley train being formed by a braided stream emerging from a glacier. Hidden Glacier, Alaska. (After Gilbert, U. S. Geological Survey.)

Kames. These are hills of stratified glacial débris with rounded outlines commonly from 50 to 150 feet high. They may exist as isolated hills or in small groups (Fig. 280), or they may be associated with unstratified deposits of moraines. When they are grouped, deep depressions occur between the hills, giving rise to what is called "knob and kettle" topography. They occur most generally in valley bottoms, but sometimes on hillsides or even on hilltops. They are rather common and widely distributed over the great glaciated region of the northern United States, particularly in association with terminal and recessional moraines. Sometimes they form so-called "kame-moraine" ridges. Kames form at the margins of glaciers by débris-laden streams which heap up the mate-

rials (usually sand and gravel) as they emerge from the ice. Sometimes the débris-charged water rises as great fountains. Kames are now actually in process of construction alongside some of the great Alaskan glaciers.

FIG. 280. Kame hills, five miles west of Gloversville, New York.

Eskers. These are long, usually winding, low ridges of stratified glacial material, mainly consisting of sand and gravel. They are seldom over 75 to 100 feet high, and their crests are generally narrow and rather even (Fig. 282). They are usually less than a mile long, but in Scandinavia and elsewhere individual eskers have been traced many miles. They often look like artificial railway embankments. They were formed by deposition in streams, choked (or overloaded) with glacial débris, either in channels on glaciers or in tunnels beneath the ice.

FIG. 281. Detail view showing the structure of a kame deposit. Gloversville, New York.

FIG. 282. **Part of an esker, showing its winding course.** North Creek, New York.

GLACIAL EFFECTS UPON RELIEF, SOILS, AND DRAINAGE

Effects upon Relief. We have already shown how valley glaciers have brought about considerable topographic changes by straightening, deepening, widening, and steepening the sides of their valleys, and also how they have developed cirque basins and "knife-edge" ridges. We have also stated that the erosion by the great ice sheets of the Ice Age did not profoundly alter the relief of the country over which they moved. Although many hills and low mountains were somewhat scoured and rounded off, and many valleys, especially those of softer rocks approximately parallel to the direction of ice movement, were appreciably deepened and widened, nevertheless the major relief features were left practically unaffected by ice erosion during the passage of the great ice sheets. We may also say that deposition of rock débris by the great glaciers left the major relief features practically unaffected. There was, however, a very appreciable tendency for glacial deposits to accumulate in the valleys during retreat of the ice, but such deposits are only minor details of the larger valleys such as the Connecticut Valley of New England, and the Hudson, Mohawk, Champlain, and St. Lawrence Valleys of New York. Viewed very broadly, the vast glaciers

(mainly due to deposition in the pre-Glacial depressions) left the country somewhat less rugged than it was just before the Ice Age.

Effects upon Soils. It is not too much to say that the passage of the great ice sheets wrought a revolutionary change in the soils of the vast glaciated area of fully 4,000,000 square miles. Residual soils very largely covered the country before the Ice Age. The movement of the ice caused such soils to be removed from their places of origin, mixed up, and transported, often for long distances. Over eastern and central Canada, where the two greatest accumulations of ice occurred (Fig. 249), ice erosion predominated over deposition, while farther south and southwest glacial deposition predominated. This explains why eastern Canada (except its southern portion) is generally characterized by bare rock ledges or only thin soils, whereas deep glacial soils generally prevail over New England, the Upper Mississippi Valley, and the southern parts of Manitoba, Saskatchewan, and Alberta in Canada.

Over extensive areas, such as the upper Mississippi Valley, the soils were made deeper and richer on the average because the glacial-drift soils are there rather uniformly deep and consist of finely ground rocks of many kinds still rich in the soluble mineral foods for plants. The pre-Glacial soils were not only thinner on numerous hillsides, but also largely depleted of the rich, soluble mineral foods for plants.

In New England and considerable parts of New York and southeastern Canada, the glacial soils are not only too sandy and gravelly to be very fertile, but also usually difficult to cultivate because of the numerous glacial boulders which they contain. These features of the soils are due to the fact that the ice passed over great areas of very hard, crystalline, igneous and metamorphic rocks which, when eroded, produced large quantities of sand, gravel, and boulders.

Effects upon Drainage. Changes of stream courses, directly resulting from the presence of the great ice sheets of the Ice Age, were numerous in many parts of the glaciated area. Some of these changes were truly far-reaching. Many pre-Glacial valleys were more or less completely filled with glacial débris, causing new streams, which came into existence after the ice melted, to follow courses very different from those of their predecessors of pre-Glacial days. In other cases streams were crowded out of their valleys by the ice itself, and forced to erode new channels elsewhere. Many such new channels have been held to since the disappearance of the ice. Only a very few examples will be mentioned. Immediately prior to the Ice Age, the combined Allegheny

and Monongahela Rivers flowed northward into the Erie Basin instead of through the Ohio Valley as at present. Ice occupancy and accumulation of much glacial débris across northwestern Pennsylvania caused the change. Rock River in northern Illinois flowed southward into the Illinois River instead of southwestward into the Mississippi as at present. The pre-Glacial Ohio River followed a devious course for fully 40 miles between Cincinnati and Lawrenceburg instead of the present short-cut between the two places, the change being due to local blockading of the old valley. The Sacandaga River of northern New York formerly flowed southward into the Mohawk River instead of turning eastward into the Hudson River as at present, blockading of its old valley by drift having been the cause of the change. We have already shown how the Missouri River was forced, by the crowding action of an ice sheet of the Ice Age, many miles westward to its present position in South Dakota.

In many cases, where streams were forced to find new channels, they have carved out picturesque gorges, usually containing waterfalls. A few examples are Niagara Gorge and Falls, and Ausable Chasm, Trenton Falls, and Watkins Glen in New York.

Most of the lakes by far, of the scores of thousands in northern North America, are direct results of the glaciation of the Ice Age. Even the Great Lakes were not in existence before the Ice Age. Some of these numerous lakes occupy rock basins which were scoured out by glacial erosion, but most of them fill basins which were formed by drift deposits blockading valleys, causing the streams in them to be locally ponded. Lakes of these kinds, as well as others, are considered at some length in Chapter XIV.

Nonglacial Ice

Ice in Soils and Rocks. Over wide areas of the higher latitude regions of the earth, water in the soil freezes during at least part of the time each year, often to a depth of from 1 to 10 feet or more. On freezing, the water in soil expands, causing movements which are of geological importance. The upward movement thus caused explains why curbstones, posts, roadbeds, etc., are often upheaved, and why boulders in the soil tend to work up to the surface. This process also causes a slow *creep* of soils and rock fragments down slopes because the expansive force of freezing lifts the rock particles or fragments at right angles to the slope, and on thawing they tend to settle vertically. By

repetition of this process much loose rock material gradually creeps downhill. On steep slopes in high mountains this expansive movement due to freezing of water in the soil or talus piles may initiate landslides. It should also be mentioned that erosion is greatly checked when the soil is frozen.

Ice in Streams. Along the shores and bottoms of many streams in cold climates ice often envelops rock fragments of various sizes. When the "break-up" comes in the spring, cakes of ice carrying such débris may float long distances before depositing their loads. The shores of the St. Lawrence show many boulders which have thus been transported.

Ice in Lakes. So-called *ice ramparts* are low ridges or walls of rock débris bordering lakes and ponds, which have been formed by the crowding action of the expanding pond or lake ice upon the shores. If the pond or lake is covered by thick ice at a temperature far below freezing, and then the temperature rises rapidly, expansion of the ice cover takes place, and loose materials in the shallow water near the shores are crowded upon the latter. This process, many times repeated, often builds up conspicuous ice ramparts.

Sea-coast Ice. In the Arctic and Antarctic regions, shallow sea-water is often frozen to a depth of a number of feet for some distance out from the shore, during the cold season. Along the shore, rocks are frozen into the bottom of the ice, and some débris accumulates on top of it from the shore. When milder weather sets in, such ice "breaks up" into what is called *floe ice*. Floe ice not only drifts away and transports much rock material, but parts of it are driven by winds and tides back and forth against the shore, eroding and grinding the rocks.

Icebergs. Icebergs are formed in high latitude regions where glaciers flow down into the sea or arms of the sea and break into fragments, both small and large, which float away. Some icebergs rise from 100 to 200 feet or more above the water; extend 1000 feet or more below the surface of the water; and cover many acres or, in the Antarctic, even several square miles. Since they are derived from glaciers, icebergs carry more or less rock material within and upon them, and this is strewn over the sea bottom, often many hundreds of miles away, as the icebergs melt.

CHAPTER X

THE WORK OF WIND

Geologic Importance of Wind Work

Introduction. Wind is an important geological agent of erosion and transportation of rock material, but, considering the earth's surface by and large, it is much less effective than running water. Most people live in humid regions where the surface of the earth is protected by vegetation, and so there is a general lack of appreciation of the work accomplished by wind. But even in humid regions wind action is by no means slight. Everyone has witnessed large clouds of dust stirred up from freshly cultivated fields during periods of dry weather in early summer or late spring. Dust of this kind is often blown for miles. By such removal of soil, young crops are often injured or ruined, and the blown soil may bury other vegetation near by. The action of wind is strikingly exhibited in humid regions along and near shores of the sea and of large lakes where sands are picked up and transported in large quantities, often to accumulate in the form of dunes.

It is in arid and semi-arid regions, however, that the wind is most effective as a geological agent. The importance of wind work becomes impressive when we realize that desert conditions prevail over about one-fifth of all the lands of the earth. In deserts weathering effects requiring moisture in the air are reduced to a minimum; stream action is relatively less important as a factor of erosion and deposition than in humid regions; and frost action, due to lack of water, is comparatively unimportant. Temperature changes in deserts are, however, exceptionally great and rapid, as between night and day, and so the rocks, which are nearly everywhere directly exposed because free from vegetation, are broken up relatively fast as a result of repeated and rapid expansion and contraction in combination with other processes.

Wind Work Classified. Wind is another significant agent of gradation. Like its companions, running water and glaciers, its gradational work is mainly twofold, (a) erosion and (b) deposition. The atmosphere as an agent of weathering has already been discussed. As a weathering agent it need not be in motion like wind. Wind *erosion*

(involving transportation) is mostly of two types—*deflation* and *corrasion.* Both, of course, accomplish removal and transportation of rock material, but the first is done without the aid of tools whereas the second requires the assistance of grinding materials.

Winds not only erode, transport, and deposit rock materials, but also stir up waves and shore currents which in turn become effective and important agents of gradation, as discussed in Chapter X.

Wind Erosion

Deflation. It is as an agent of deflation that wind accomplishes its greatest amount of geologic work. Already loosened particles of rock, such as dust and sand, are picked up and carried from one place to another. Such activity may be extremely conspicuous in regions where vegetation is sparse or lacking.

What are some of the sources of the finely divided rock material which is removed and transported by winds? Most of the material by far is picked up from dry surfaces of loose, fine materials of all kinds in all sorts of regions, but especially in deserts where such materials are blown about by every wind. The broad playas of desert basins are prolific sources of large quantities of wind-blown material. Some wind-blown material is derived directly from rock ledges by the erosive action of the wind itself, as explained beyond. Considerable quantities of dust are contributed to the atmosphere by explosive eruptions of volcanoes whereby lava is pulverized and shot far into the air (Fig. 139). During the explosions of Krakatoa volcano, a tremendous quantity of finely divided and pulverized rock was forced miles into the air, and some of it was carried completely around the earth and remained suspended for many days, causing the famous red sunsets of 1883. Several cubic miles of volcanic dust were forced out of Katmai Volcano, Alaska, in 1912 when the mountain exploded. This dust caused darkness for many miles around for more than two days; and at a distance of 100 miles it accumulated to a depth of 10 inches.

The almost inconceivable transporting power of strong winds over deserts is illustrated by the well-known "sand storms" of the Sahara Desert. In such a great storm many cubic miles of dust and sand-laden air sweep miles across the country. It has been estimated that one cubic mile of air in such a storm carries at least 100,000 tons of rock material. Dust from the Sahara is known to be carried hundreds of miles out into the Atlantic Ocean. According to an estimate, a great

storm in 1901 carried nearly 2,000,000 tons of finely divided rock material from northern Africa into Europe. In two days some of the dust fell in Italy and in three days some of it reached central Germany.

Wind Corrasion. Wind picks up and carries along great quantities of dry, loose, finely divided material, but of itself it has little or no power to abrade or corrade solid rocks. Wind, like water, effectively scours rocks when properly supplied with tools, that is, when it has

FIG. 283. An approaching dust storm in Union County, New Mexico, May 21, 1937. (Photo by Al Carter for U. S. Soil Conservation Service.)

rock fragments with which to work. When fine material, especially grains of sand, are driven by wind with high velocity against barren rocks, the latter are worn and often polished by the process. The principle involved is that of the sand blast used in cleaning and polishing decorative and building stones, and in etching glass. Where rock ledges show many local variations in composition and hardness, they are often etched by wind and rain work into very irregular, and often fantastic, forms.

A surprising amount of corrasion may be accomplished by the wind, under very favorable conditions, in a short time. A plate glass window

in a Cape Cod lighthouse is said to have been worn to opaqueness during a single hard wind storm. Window glass directly exposed to hard winds on Cape Cod is known to have been completely worn through within a few weeks or months.

Automobiles driven in violent dust storms in the desert, and even elsewhere, not infrequently have had their windshields rendered useless, and the paint more or less removed from them, by wind-driven sand.

FIG. 284. A striking example of rock forms scoured and sculptured by wind corrasion. At times loose sand is picked up from the bottom and driven at high velocity through this troughlike depression. Rogers Dry Lake, Mohave Desert, California. (Photo by E. Blackwelder.)

Wind-driven sand has its greatest erosive power relatively close to the ground because the heavier and larger fragments, not being lifted so high, there accomplish the greatest work. Telegraph poles in desert regions often must be especially protected else they will be cut down by sand driven against their bases. Pebbles and boulders on deserts sometimes have more or less angular faces carved upon them by wind erosion, and rock platforms are often kept worn smooth and hard. The finer products of rock weathering are often removed about as fast as they form. Larger rock fragments on the surface are gradually worn smaller. In the arid and semi-arid southwestern states of the United States, cliffs are often undercut by the corrasive action of the wind, sometimes with the development of large caverns. The famous Sphinx,

and even the pyramids, of Egypt have been very considerably roughened by the action of wind.

Wind and Stream Erosion Compared. Careful observations during the last ten to twenty years have led to the conclusion that the corrasive action of wind is much less generally effective in sculpturing and cutting away rocks than formerly surmised. There is, however, no doubt about the local efficacy of wind corrasion when conditions are very favorable (Fig. 284).

Stream erosion, even in desert regions, accomplishes much more work than wind erosion because when it rains it is often in the form of so-called "cloudbursts." In desert areas of high relief, therefore, torrents of water rush down the stream courses carrying heavy loads of rock débris derived from the abundant weathered rock material almost unprotected by vegetation. The deep sculpturing of the desert mountains of Nevada, Arizona, western Utah, and eastern California is very largely due to water, and not wind, erosion.

The writer has observed thousands of well-exposed outcrops in the vast western desert showing plain evidence of mechanical and chemical weathering, but with little or no evidence of wind corrasion.

The greatest geologic work accomplished by the wind seems to be deflation, the picking up and transporting of really large quantities of finer materials which either were laid down through the agency of running water or were produced by weathering.

Wind Deposition

Dunes. Hills of wind-blown sand are called *dunes*. They are formed in much the same manner as snowdrifts. They are abundant in many regions, as for example along the middle Atlantic Coast of the United States; around the southern end of Lake Michigan in Dune Park, Indiana; and in the desert portions of the western United States. Dunes mostly form in deserts; on and near sandy shores of lakes or oceans where the wind blows toward the land; and on and near river flood plains, especially in arid regions where the volume of water varies greatly. Dunes seldom attain heights greater than a few hundred feet, although some in the Sahara Desert are said to be more than 1000 feet high.

A dune may begin to build up where there is a slight irregularity of surface or some obstacle, such as a boulder, causing a local check in the velocity of the wind with resultant deposition of some of the load it

carries. Once the pile has started, its growth is accelerated by its own shape and size. Where the direction of the wind remains fairly constant, the tendency is for a gentle slope to develop on the windward side, and a steep slope on the lee side. Sand blown up to the crest of the windward side is caught in a relative calm with a back-eddy on the lee side of the hill, and there deposited. The lee side is steeper because the

FIG. 285. A crescentic sand dune (or barchan) in Wyoming. (Photo by E. E. Smith, U. S. Geological Survey.)

FIG. 286. Sharp-crested sand dunes near Stovepipe Wells in Death Valley, California. Note the numerous furrows produced by tiny sand slides at the lower right. (Photo by Dick Freeman, Los Angeles.)

sand rolls or slides down its slope (Fig. 286). Smaller dunes are often somewhat crescent-shaped, caused by the wind driving sand both over and around the dune (Fig. 285). This type is known as a *barchan.* When the winds are rather variable in direction, the sand dunes are more irregular in shape.

Fig. 287. A detail view showing a striking exhibition of cross-bedding produced by wind action many millions of years ago. The rock is now hard sandstone exposed by erosion. Pine Canyon, tributary to Zion Canyon, Utah.

Dune sand is usually crudely stratified, with prominent cross-bedding, due to the varying velocity of the wind which causes alternately larger and smaller sand particles to be driven up the slopes and deposited in layers (Fig. 287). Sand dunes are often beautifully ripple-marked on their surfaces by more or less parallel ridges an inch or more high. (Fig. 288).

Migration of Dunes. Unless prevented by vegetation, dunes usually migrate in the direction of the prevailing wind. The migration is caused by the blowing of the sand up the windward slope, and its deposition on the steeper leeward slope. On a dry, windy day the sand can be seen blowing over the crest of a dune. The rate of migration is of course determined by several factors. Most dunes migrate at rates of from a few feet to more than 100 feet per year. A case of unusually rapid movement was that at Kunzen on the Baltic Coast where a large dune encroached upon, buried, and then uncovered a church, migrating about eight miles between the years 1809 and 1869. A pine forest, covering hundreds of acres on the coast of Prussia, was destroyed by

migration of dunes between 1804 and 1827. In many places portions of farms have been ruined by migration of dunes within the lifetime of

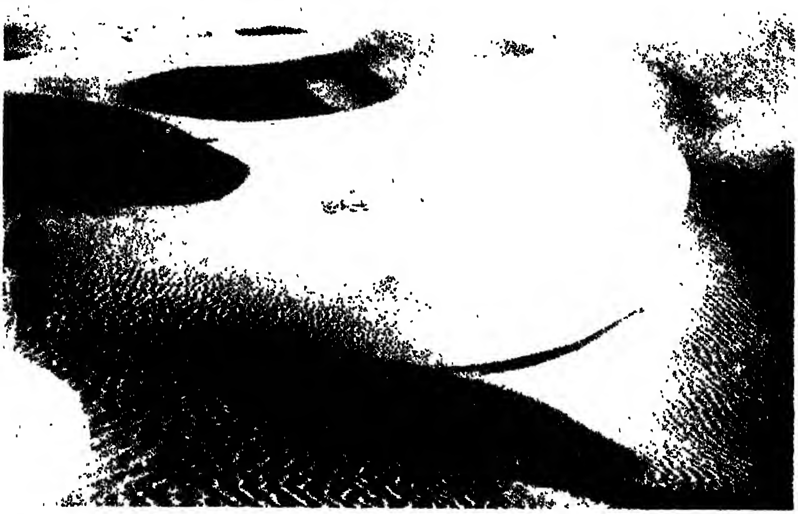

FIG. 288. A group of ripple-marked sand dunes in the Imperial Valley, California. These are in the extensive dune area shown in Fig. 293.

single owners. Forest trees have, in many places, been buried, killed, and then uncovered by drifting sand (Fig. 290). Such phenomena are well exhibited in Dune Park, Indiana.

FIG. 289. A sand dune advancing upon a forest. Near Port Burwell, Ontario.

A region of special interest, which has been carefully studied by Reclus, is the shore of the Bay of Biscay northward from Bayonne. "The sea here throws every year upon the beach, along a line 100 miles in length, some 5,000,000 cubic yards of sand. The prevailing westerly winds, continually picking up the surface particles from the seaward side,

whirl them over to the inland or leeward slope, where they are again deposited, and the entire ridge by this means alone moves gradually inland. In the course of years there have thus been formed a complex series of dunes, all approximately parallel with the coast and with one another, and of all altitudes up to 250 feet. These are still marching steadily inward, though at the rate of but three to six feet annually, and whole villages have more than once been torn down to prevent burial, and rebuilt at a distance, to be again removed, within 200 years" (G. P. Merrill).

FIG. 290. Tree trunks being uncovered by a migrating sand dune. Near Port Burwell, Ontario.

Deposition over General Areas. Wind-blown material does not, by any means, always accumulate in the form of dunes and ridges. Wind

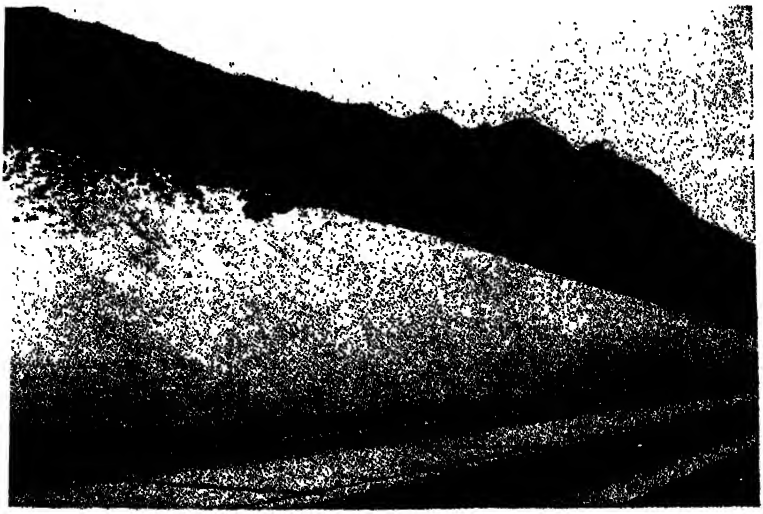

FIG. 291. Wind-blown sand from a near-by seabeach (to right of picture) piled 100 feet high against a steep hillside. Coast highway, 12 miles southeast of Oxnard, California.

work often tends to level off large areas by removing loose materials from higher lands and depositing them in intervening depressions, or piling them against bases of mountains. This is true on a grand scale in portions of the Sahara Desert where certain wide areas of bedrock are kept free from sand by wind erosion, and the sand is piled against the mountain bases, and even up the slopes to heights of 1000 to 2000 feet. In the Great Basin region of the Western United States somewhat similar phenomena are not uncommon.

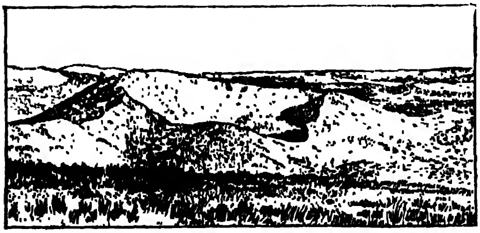

FIG. 292. Sand dune with a "blow-out" in its top. Near Julesburg, Colorado. (After U. S. Geological Survey.)

The ruins of the once great cities of Nineveh and Babylon are largely buried under wind-blown sand and dust. Evidence has been presented to show that the climate of western and central Asia is now considerably drier than it was a few thousand years ago.

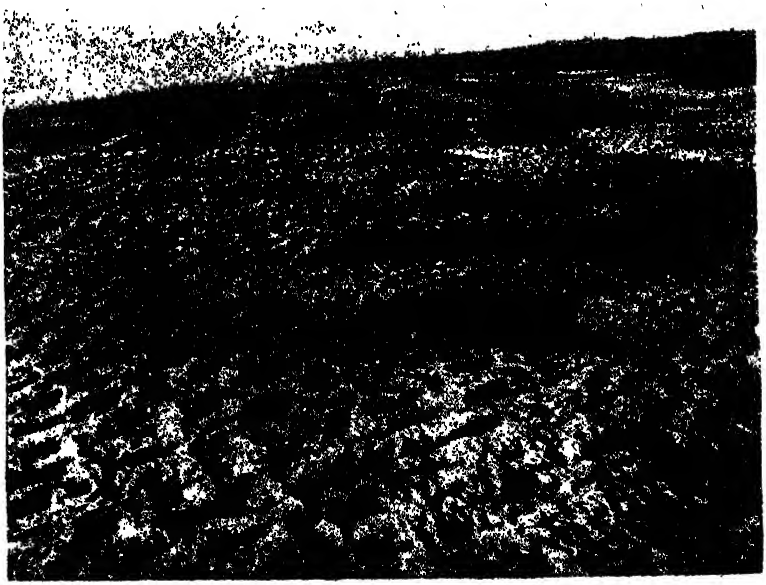

FIG. 293. An aerial view of part of the great sand dune region in the eastern part of Imperial County, California. (© Spence Air Photos.)

Loess. A kind of deposit of special interest, which is mainly or partly of wind-blown origin, is called *loess*. It is usually a fine-grained, unstratified, yellow to brown loam or silt which, though very slightly consolidated, has the remarkable property of standing in the form of high, very steep slopes or cliffs where it has been cut into by erosion (Fig. 294). It sometimes contains shells of land animals. It forms extensive deposits, commonly from 10 to several hundred feet thick, in various regions.

FIG. 294. A roadway through a deposit of loess in China. Note the vertical structure. (Photo by Bailey Willis for the Carnegie Institution of Washington.)

Certain valleys of northern Europe, especially the Rhine, contain loess deposits. Extensive deposits occur in Argentina. Thousands of square miles in Iowa, Nebraska, and Kansas (particularly in the Missouri and Mississippi Valleys) are covered with loess which is seldom more than 100 feet thick. This is believed to represent the fine, loose material blown by the wind from the adjacent regions just after the withdrawal of one of the great glaciers of the Ice Age. The loose glacial soils were then protected by little or no vegetation. Many thousands of square miles of northern China are covered with loess, much of which may have been blown from the Mongolian desert. It covers both mountainsides and valleys to depths probably as great as 500 feet. Some of this loess has probably been reworked and deposited by water.

CHAPTER XI

THE SEA AND ITS WORK

Introduction

According to good general usage the terms *sea* and *ocean* are practically synonymous and refer to the whole continuous body of salt water, including its numerous embayments, which covers a large part of the earth's surface. The term *sea* is preferred in this book. Such names as the "Sea of Galilee" or the "Dead Sea" are misnomers because, being inland bodies of water (one salt and the other fresh) not connected with the great general body of salt water, they are really lakes.

Certain scientific distinctions of use to oceanographers, geographers, and geologists, may be made as follows. The *oceans* are very large bodies of deep-sea water occupying basins between continents. Thus we have the Atlantic, Pacific, Indian, Arctic, and Antarctic Oceans whose average depth is about two and a half miles.

Epicontinental seas, seldom over 600 feet deep, occupy the narrow platforms (continental shelves) which border the lands nearly everywhere. They are, in other words, shallow-water, landward extensions of the open ocean. They are also known as *shelf* or *marginal seas.* A fine example is the sea covering the continental shelf off the eastern United States. Other examples, less open to the ocean, are the North Sea, the Yellow Sea, and the Gulf of St. Lawrence.

Epeiric seas are also shallow (seldom over 600 feet), but they lie well within continental regions and their connection with the ocean is less open. Probably the finest example is Hudson Bay. The Baltic Sea, extending northward into the Gulf of Bothnia, is another good example. During the geologic ages, particularly the Paleozoic era, it was epeiric seas which repeatedly spread over, and withdrew from, large and small parts of continents. Most of the exposed marine strata of North America and Europe were deposited in epeiric seas.

A *mediterranean* is a special type of sea much like an epeiric sea, but it has depths of thousands of feet. The Mediterranean Sea is the finest example, and the Caribbean Sea may be classed in this category.

Geological Importance of the Sea

The deep seas, such as the Atlantic and Pacific Oceans, are geologically very old, and they have probably remained in essentially the same positions for hundreds of millions of years. Shallow seas have, however, as already mentioned, spread over, and disappeared from, large and small parts of continents time and again during the geologic eons. It is plain, as shown by the character and origin of the marine strata now exposed on the lands, that no deep sea (true ocean) ever spread over any considerable part of a submerged continent.

The sea is now, and has been through known geologic time, the greatest theater of sedimentation. Shallow-water marine strata of practically all known ages are very extensively exposed within the continents. Such strata have been piled up to a total thickness often reaching 5 to 15 miles. In many cases they have been greatly disturbed (folded and faulted) out of their original position and deeply eroded, thus exposing to view the records which they contain. Had it not been for the accumulation of these marine strata, far less would be known about many of the great and small physical changes through which the earth has passed.

Marine strata also contain countless myriads of remains and impressions of animals and plants (i.e., fossils), and thus we have a very important key to a knowledge of the kinds, distribution, and history and evolution of life on the earth through hundreds of millions of years.

The sea has, through the long ages, been incessantly at work cutting into and modifying many parts of the bordering lands.

The climatic influence of the sea has also been of real importance. Thus the moisture in the air, rain which forms streams, and snow which forms glaciers, all have their sources very largely in the ocean, and these agents in turn accomplish great work of weathering and erosion.

Extent and Depth of the Sea

It is well known that the waters of the sea cover nearly three-fourths of the surface of the earth. The sea is about 45 times as large as the United States, that is, it covers approximately 140,000,000 square miles. The average depth of the oceans of the earth is about two and a half miles. If the sea were present universally, everywhere with the same depth, it would be almost two miles deep. Yet this vast body of water

is an extremely thin layer when compared to the earth's diameter of nearly 8000 miles.

The Pacific is the deepest of the oceans, its average depth being about two and three-quarters miles. The deepest sounding ever made was 35,410 feet, or more than six miles, off the southern Philippine Islands. It is known as the *Mindanao deep.* The second greatest depth known is 34,623 feet, in the *Tuscarora deep,* off southern Japan. There are many places in the Pacific Ocean where the water is four to five miles deep. The deepest sounding ever made in the Atlantic Ocean was nearly 31,000 feet, off Puerto Rico.

Composition of the Sea Water

Many substances are known to be in solution in sea water, but, in spite of this, the composition is remarkably uniform. The most abundant substance by far in solution is common salt. In every 100 pounds of sea water there are three and one-half pounds of mineral matter of various kinds dissolved. Nearly 78 per cent of the dissolved matter is common salt. The other principal constituents in solution are chloride and sulphate of magnesia, and the sulphates of lime and potash. All other dissolved mineral substances together make up less than one per cent of the total. It has been estimated that if all the dissolved mineral matter in the sea could be brought together it would form a layer 175 feet thick over the whole sea bottom. The salts of the sea have been mostly supplied by the rivers which in turn have derived them from the disintegration and chemical decay of the rocks.

In addition to the salts in solution, there are certain gases, chiefly atmospheric, that is, they have been dissolved mostly from the air. Organisms and submarine volcanoes supply some gases. The principal gases in solution are nitrogen, oxygen, and carbonic acid gas.

Temperature of the Sea

The temperature of the surface sea-water in the torrid zone is from 75° to 80°. From this there is a fairly gradual decrease to about 28° in the polar regions. The freezing point of sea water is 28° instead of 32° as for fresh water. Many variations in the temperature of the surface sea-waters are due to ocean currents.

In the torrid and temperate zones, at depths greater than 4000 to 5000 feet, the temperature of the sea is always lower than 40°. Fully

two-thirds of the ocean water is, therefore, colder than 40°, and, generally speaking, the water grows colder with increasing depth, even reaching 31° at great depths. At the greatest depths under the equator, the water is not far from the freezing point. In the polar regions the sea is of course, from surface to bottom, nearly everywhere at or near the freezing point.

Life in the Sea

Both plants and animals exist in countless numbers in the sea. The animals range in size from single-celled microscopic forms to that of the whale. Lower orders of animals are much more common than the higher. Among the higher forms of animal life are whales, seals, walruses, turtles, and untold millions of fishes. The plants are mostly simple forms, like seaweeds, which range in size from microscopic to several hundred feet long.

In the upper few hundred feet of the sea, and more especially at and close to the surface, vast swarms of organisms, both plant and animal, exist. Most of these are tiny to microscopic in size. Since plants depend upon light for their existence, they are confined wholly to the upper few hundred feet of the sea.

Murray, the great student of the ocean, says: "We know that the whole of the surface waters of the ocean are crowded with minute unicellular algae (seaweeds, etc.) which are ever busy, under the influence of sunlight and chlorophyll, converting the inorganic substances in the sea water into organic compounds, which in turn supply not only the food of the vast majority of marine animals which live in the surface and intermediate waters, but also of the myriads of creatures living near and on the sea floor, miles beneath the level to which the sun's rays can penetrate. . . . The bodies of the minute unicellular algae, which often have calcareous, siliceous, or chitinous shells, fall to the bottom after death, together with the dead bodies of the animals which browse in the meadows; accumulating on the surfaces of the deep-sea oozes and clays, they supply nourishment for the creatures that crawl on the bottom of the sea."

Besides the minute forms there are "many larger floating and swimming species, and some clinging and fixed forms attached to floating bodies such as logs and seaweeds, or to swimming animals. In the (so-called) Sargasso Sea, for example, there is a miniature world of plant life and dependent swimming, crawling, and fixed forms of animal life. Among the larger animals are numerous fishes, some like the

herring and mackerel, swimming in great schools, others moving singly like the shark and swordfish. The whale also roams in the surface and upper layers of the ocean. A multitude of floating species of jellyfish and other forms of animal life also inhabit this zone" (Tarr and Martin).

Compared to the countless myriads of organisms in the upper layer of the sea, the animal life of the great bulk of ocean water is much less profuse. Deep-sea investigations have, however, revealed the existence of many types of animals. Many of these are very curious-looking creatures with no counterparts in the land waters. Some of them, like the so-called "sea-lily," belong to types of geologically old animal forms as proved by the remains of similar creatures embedded in the rocks of the earth. Prominent among the swimming creatures of the deep sea are certain strange-looking fishes. Other types of animals either crawl on the sea-bottom or burrow through the soft, oozy substances there.

Along the seashore, and in the shallow waters near shore, especially in the warmer zones, animal life is abundant and varied, including many shelled forms, such as clams, mussels, and barnacles; crustaceans, such as lobsters and crabs; so-called starfishes of various sorts; and, in clear waters not colder than 68°, corals. In many places, as for example around Florida, and just north of both Cuba and Australia, the corals have built up great reefs by the accumulations of the carbonate of lime skeletons which these tiny animals have secreted from the sea water.

Tides

The tide is a great wave hundreds of miles across, and only a few feet high in the open ocean. It is caused chiefly by the attraction of the moon and, to a slighter degree, of the sun. The highest wave results when the sun and moon pull together along the same line. Since there is a reaction about equal to the action on opposite sides of the earth, two tidal waves are produced at the same time. Because of the earth's rotation, were there no lands to interfere, each of these great, low waves would pass around the earth every 24 hours, or, in other words, one would pass a given place every 12 hours. As it is, a tidal wave strikes the shore of each hemisphere every 12 hours, the water piling up on the shore until the crest of the wave comes, after which the water recedes gradually. In bays and estuaries, especially those which are V-shaped with their wide portions toward the ocean, there is a tendency for the

water to pile up unusually high. A very remarkable example is the Bay of Fundy, in which the tidal wave often rises 30 to 50 feet.

Ocean Currents

In the equatorial regions of both the Atlantic and Pacific Oceans, the steady friction of the persistent trade winds produces wide, westward moving, surface currents. Each of these currents, on striking the continental coast, divides into two portions, one moving southward and the other northward. Each portion then crosses the ocean eastward and finally turns back into the equatorial belt. Thus there are in each great ocean two vast eddies, one north, and the other south, of the equator, with relatively quiet water in the midst of each. When a wide, slow current aproaches the land, its water tends to accumulate and, where the slope of the land is favorable, it becomes narrower and swifter, giving rise to so-called *streams.*

The *Gulf Stream* is formed by a crowding of part of the deflected equatorial current of the Atlantic into the Caribbean Sea and the Gulf of Mexico. Where it emerges from the narrow strait between Florida and Cuba, it has a speed of 100 miles per day. As it moves along the eastern side of the United States, and into the Atlantic Ocean, it becomes gradually much wider, and its velocity is reduced to about 10 miles per day.

Topography of the Sea Floor

If we make a general comparison with the surface of the land, the bottom of the sea, well out from the land, is generally the smoother. Little of the sea bottom compares with the ruggedness of the mountains, and even the more level portions of the land surface nearly always show many sharp, minor irregularities, such as stream trenches; but the sea bottom is characterized more by its smoothness of surface. Under the sea there are, however, mountain-like ridges, plateaus, submarine volcanoes, and valleys, the deeper valleys being known as *deeps,* but such features, 100 miles or more from the shore, seldom show roughness of relief such as characterizes similar features on land.

One of the most remarkable relief features of the ocean bottom is known as the *continental shelf* (Fig. 307). It is a relatively narrow platform covered by shallow water bordering nearly all the important lands of the earth. Usually the water increases in depth seaward over this platform, but it seldom reaches a depth of more than 600 feet. The

continental shelves of the world cover about 10,000,000 square miles, or about one-fourteenth of the area of the sea floor. "The break in the slope at the outer margin of the continental shelves is found at depths of approximately 400 feet off most of the coasts of the world. It seems probable that this uniformity was developed by the cutting of the waves at a time when the sea level was much lower than at the present time. Such a lowering did occur at the time when the sea water was extracted from the ocean and piled up on parts of the land to form the great continental glaciers of the past" (Shepard).

On the way from New York to Europe, a ship sails over the continental shelf for about 100 miles, the water gradually increasing in depth to about 600 feet. Then there is a comparatively steep descent (called the *continental slope*) into the great *ocean abyss* which is two to three miles deep (Fig. 307). The floor of this abyss, stretching across the Atlantic almost to the shores of Europe, is generally lacking in sharp topographic changes. A little more than half way across, the ocean bottom rises as a kind of submarine plateau or ridge a few hundred miles wide, with water not more than one to two miles deep over it. This submarine ridge runs roughly north and south with a winding course through nearly the whole Atlantic Ocean. Within a few hundred miles of Europe the sea bottom begins to rise on a continental slope to a continental shelf which, with its shallow water, extends to the shore.

Waves

There are several kinds of sea waves, such as wind waves, tidal waves, tsunamis, etc. Our present concern is chiefly with wind waves which are produced by the friction of wind blowing over the sea surface. Because important work of both erosion and deposition is accomplished by waves, certain facts in regard to them will be mentioned.

It is an interesting fact that, in calm weather, the form of a wave advances rapidly, but the water itself advances at a much slower rate. This is because the wave form is caused by particles of water moving in vertical, nearly circular orbits. A familiar demonstration of this fact is the rising and falling, and to-and-fro movement, of a chip when small waves pass under it on a pond. The chip moves forward very slowly unless wind pushes it along.

When a wave moves shoreward into shallow water on a gently sloping bottom it becomes shorter and higher, and the lower part has its motion gradually checked so that the unimpeded upper part rushes

over it, thus forming a *breaker*. This in turn causes a rush of water still farther forward to the shore as may be seen on a beach. This forward rush of water usually carries much sediment with it. After the water thus moves upon a sloping shore, it returns seaward, in part down the slope, as an undercurrent called *undertow,* and in part as *rip currents*. Where undertow meets incoming waves much turbulence is produced, and its seaward motion is stopped.

Rip currents, according to Shepard, "are masses of water that move straight out from the beach carrying water at both the surface and near bottom, but chiefly at the surface." According to Grant, rip currents, which occur more or less locally at varying times and positions, and seldom extend out more than a few hundreds of feet, are much more effective than undertow in carrying sediment seaward.

If a wave strikes the shore obliquely, the returning water moves down the slope at about right angles to the shore, and the next wave drives part of this same water against a slightly different part of the shore. Repetitions of this process cause the development of a *shore current* whose strength may be augmented by the prevailing wind. These shore currents are, as pointed out beyond, important in the building of certain shore forms.

Storm waves on the sea are commonly from 15 to 20 feet high and several hundred feet long, but they may at times be twice that size. Such waves attain velocities of from 20 to 60 miles per hour.

In considering the work done by waves, it is important to know that large waves will move coarse sand and gravel at depths of from 50 to 100 feet, and fine sand at a depth of several hundred feet.

The force with which waves strike bold, rocky shores is also an important consideration because of its influence in wave erosion (Fig. 296). The force of impact of waves is ordinarily from 1000 to 2000 pounds per square foot, but in great storms it is several times as much when blocks of rock many tons in weight are moved.

MARINE EROSION

How Waves Erode. When a wave dashes against a rocky shore or cliff, water is forced into many cracks and other openings, causing a hydraulic pressure which tends to disrupt blocks of rock in the face of the cliff. Also many fissures and crevices are suddenly filled with compressed air which, on retreat of the wave, has its pressure relieved quickly, thus producing an explosive recoil which often dislodges masses

of rock. The very impact of the wave against a shore may be sufficient to force off rock material from a cliff, not only of soft or loose rock, but also of hard rock if there are masses of it already sufficiently loosened by jointing or weathering. A minor factor in the removal of rock material by waves is the solvent action of sea water. This is particularly true of limy rocks where organisms increase the acidity of the water and hence its solvent action.

Waves are most effective as agents of erosion through their grinding action. Waves, like running water, wind, and glaciers, erode most effectively when properly supplied with rock fragments as tools with which to work. When strong waves, armed with rock fragments already dislodged from the shore, repeatedly strike a rocky shore or

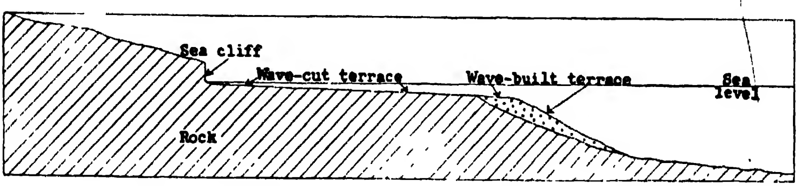

Fig. 295. Diagrammatic structure section illustrating the development of the sea cliff, wave-cut terrace, and wave-built terrace.

cliff, they become powerful agents of shore destruction. Such battering of the rock fragments of all sizes, and their rubbing against one another when carried by the undertow, soon cause the fragments to become rounded (Fig. 297). The older fragments are worn down from pebbles and boulders to fine sand and mud, while new fragments are being derived from the shore. Some corrasive effect is also produced by waves shifting sediment on the bottom where the water reaches depths to several hundred feet.

Sea Cliff and Wave-cut Terrace. Where waves are at work cutting into a shore of at least moderately high-land, a steep front facing the sea soon develops. This is called the *sea cliff*. At first the waves may attack the whole face of the cliff, but after a time the cliff becomes so high that the waves attack only its lower portion. By this undercutting process, aided by weathering, the material from the higher portion of the cliff breaks away and falls to the base to furnish more tools with which the waves may batter the cliff.

As the sea cliff retreats, a shallow-water shelf called the *wave-cut terrace* (Fig. 295) develops, over which the water increases in depth

seaward to the limit of wave action, that is, to a depth of hundreds of feet. Such a terrace will not, as a rule, be cut many miles wide, because the waves, in moving over the shallow-water shelf, lose their power gradually on account of friction on the bottom.

Much material cut away and ground up by the waves is carried seaward usually to build up the *wave-built terrace* (Fig. 295), which is something like a submarine delta. Some of it is carried by shore currents to form spits, bars, etc., as explained beyond. Observations during

Fig. 296. Sea waves eroding a rocky coast. Santa Cruz, California.

the last decade have led to the conclusion that large, well-defined wave-built terraces are by no means as common as was once supposed. This matter is discussed more fully beyond under "Marine Deposits."

Rate of Retreat of Sea Cliffs. The rate of retreat of sea cliffs is known in many places. The rate is, of course, dependent upon various factors, particularly the force and persistence of wave action and the nature of the rocks attacked. A cliff of loose material is cut back often so rapidly as to be a matter of common knowledge. A remarkable example is the island of Heligoland on which, until recently, was located the powerful German fort, guarding the Kiel Canal. In the year A.D. 800 this island had 120 miles of shoreline; in 1300 it had 45 miles of shore; in 1649 only eight miles; and in 1900 only three miles.

In southeastern England "whole farms and villages have been washed away in the last few centuries, the sea cliffs retreating from

7 to 15 feet a year." A church located a mile from the seashore near the mouth of the Thames in the 16th century now stands on a cliff overlooking the sea.

An island of soft-rock material in Chesapeake Bay covered over 400 acres in 1848, and the waves reduced it to about 50 acres by 1910.

FIG. 297. Sea cliff with wave-worn boulders at its base. The picture was taken when wave action was relatively slight. Near San Pedro, California.

Certain cliffs of soft material on the island of Martha's Vineyard retreated five and one-half feet per year between 1846 and 1886. Wave erosion on very hard rock, like granite, is far less rapid.

Sea Caves, Coves, Stacks, and Arches. Many irregularities often develop during the retreat of the sea cliff and the cutting of the wave-cut terrace. *Sea caves* are often produced along the bases of cliffs by wave action, especially where masses of weaker rock lie at or near sea level.

If, along a coast, masses of more easily eroded rocks are separated by harder or more difficultly eroded rocks, the waves will cut the former back faster to form *sea coves,* while the latter project into the sea to form *headlands* (Fig. 303).

If part of the roof of a sea cave collapses, or if two caves on

FIG. 298. Horizontal strata being irregularly cut into by sea waves. Note the sea caves, coves, and headlands. Near Laguna Beach, California.

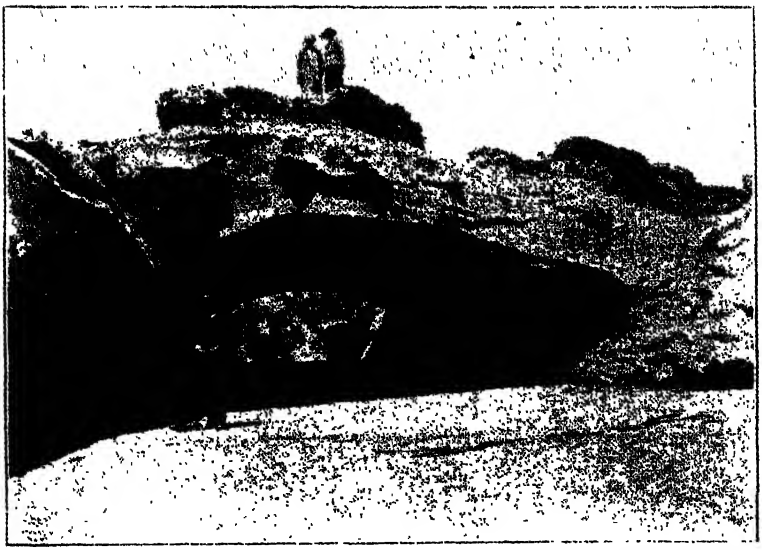

FIG. 299. A natural bridge (sea arch) carved out by high-tide waves. Santa Cruz, California.

opposite sides of a sharp headland unite, a *sea arch* results (Fig. 299). The waves will continue to batter the arch until it collapses.

Unequal wave erosion along a rocky coast often leaves isolated portions of cliffs known as *stacks* (Fig. 300). They are at most very temporary features. A famous example is the Old Man of Hoy in the Orkney Islands. It is an isolated joint column of colored sandstone 600 feet high. Many examples of stacks occur on the New England coast, and on the Pacific coast of North America.

Plains of Marine Erosion. With sufficient uplift of land relative to sea level, a wave-cut terrace becomes a *plain* (or *terrace*) *of marine erosion*. The surface of such a plain, like that of a stream-developed

FIG. 300. Remnants of wave erosion (so-called "stacks"). La Jolla, California.

peneplain, cuts across all kinds of rocks irrespective of their composition and structure. On a peneplain, however, the rock waste is characteristically a residual soil, while that of a plain of marine erosion consists of water-worn transported material rather uniformly spread over the surface. The surface of a newly upraised marine plain is usually smoother than that of a plain of stream erosion, and the erosion remnants left by the waves are steeper sided than those on a peneplain formed by streams. A remarkable example of a long plain of marine erosion with steep-sided, isolated masses (former islands) rising above its surface occurs on the eastern side of India.

A conspicuous marine terrace, usually with an altitude of approximately 100 feet, faces the sea at many places along several hundred miles of the coast of southern California (Fig. 301). Still higher terraces, proving successive uplifts of the sea floor, are also well preserved along parts of this coast.

Southern New England, in general, slopes from Massachusetts southward to tide water, but it does so by a series of more or less well-defined, broad, steplike platforms. It has been advocated recently that these platforms represent marine plains or terraces raised out of the sea

FIG. 301. A marine terrace (with "stack") being cut into by sea waves. Near Pismo, California.

by successive movements in relatively recent geological time. In this region the terraced surface, being older, has been much more modified by weathering and erosion than that around Los Angeles.

Before leaving the consideration of marine erosion, mention should be made of the fact that, during the long eons of earth history, the sea has been a much less important factor in cutting away the lands than the subaërial agents of erosion, particularly streams. Subaërial agents of weathering and erosion operate incessantly over most of the land surfaces, while marine erosion is restricted to the margins of the lands, and, as a matter of fact, only to parts of the land margins because, in many places, marine deposition, rather than erosion, is taking place along coasts.

Marine Deposits

Viewed in a broad way, there are two great classes of marine deposits: (1) those laid down in shallow water comparatively near the borders of the land, that is, on the continental shelf and continental slope; and (2) the abysmal (deep sea) deposits laid down on the floor of the deep ocean (Fig. 307).

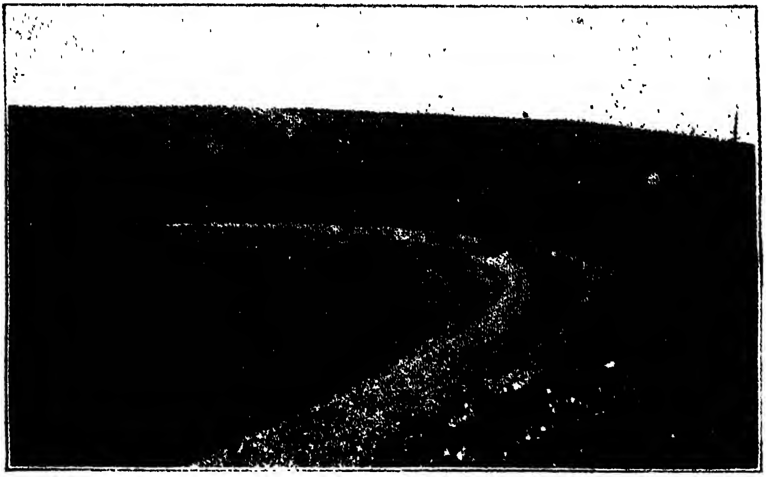

Fig. 302. A crescentic gravel beach. Conception Bay, Newfoundland. (Photo by C. D. Walcott for U. S. Geological Survey.)

Shallow-water Deposits. General statement. Marine sediments which accumulate along and near the continental borders are largely land-derived materials, that is, they are mostly sediments carried by streams from the land into the sea and, to a less extent, rock materials broken up by the waves along many shores. Practically all land-derived material is deposited within 100 to 300 miles of the shore. The quantity of such sediment carried into the sea each year is tremendous.

The continental border deposits near shore are often coarser material, such as sands and gravels, whereas farther out they often become finer material, such as muds. In many cases, however, these deposits are extremely variable because of the stirring and shifting action of waves and currents in the shallow border waters. These deposits usually contain more or less organic material, especially shells and skeletons of organisms. In some cases such organic remains predominate or even

exist to the exclusion of nearly all other material, as is true of the coral deposits (or reefs) which form only in warm, shallow water.

Materials which accumulate on the shallow-sea bottom around the borders of the lands are of great significance to the geologist because just such marine deposits, now consolidated into sandstones, conglomerates, shales, and limestones, are so widely exposed over the various continents. A knowledge of the conditions under which shallow-sea deposits are now forming is, therefore, of much value in interpreting

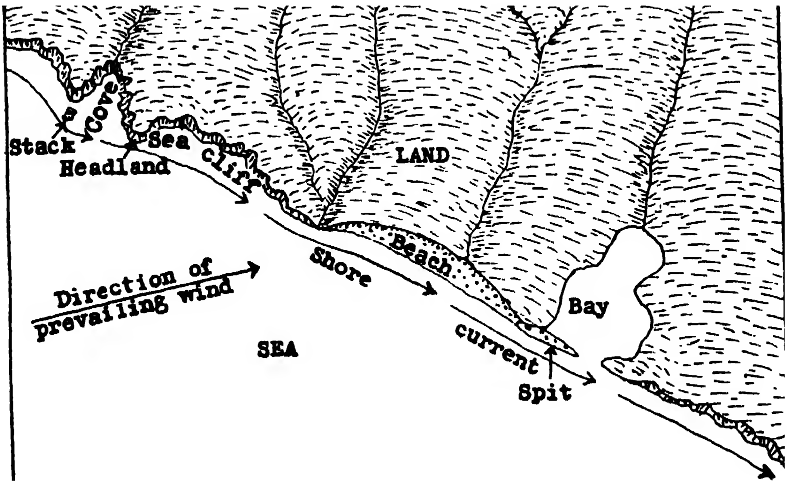

FIG. 303. Sketch map showing locations and development of shore current, sea cliff, headland, stack, cove, beach, and spit.

events of earth history as they are recorded in similar rocks which have been accumulating through millions of years of time, and which have, from time to time, been raised into land, and more or less eroded.

During recent years great advances have been made in the study of marine sediments. F. P. Shepard, a special student of this subject, has kindly furnished the writer with the following statement: "It was supposed formerly that the continental shelves were formed as a combination of the cutting back of the waves against the shore and of the building of great detrital (wave-built) terraces beyond the wave-cut shelf. It was also thought that the sediments of the shelves graded outward from coarse gravels near shore to silts and clays on the outer margins. Compilation of information gathered over many years by the

oceanographic surveys of various parts of the world gives little to confirm these suppositions. It has been found that the sediments are in general as coarse grained on the outside of the shelves as inside, and in many places coarser. Also rock bottom is reported from the outer shelf in innumerable localities instead of the sediment which was thought

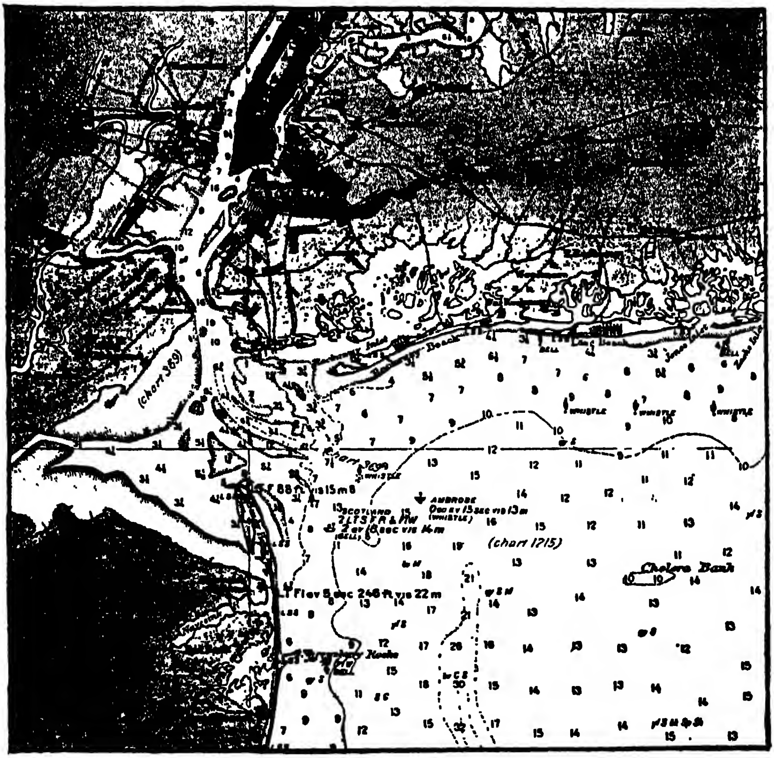

FIG. 304. Map showing beaches and spits near New York City in 1916. Figures represent depths in fathoms below low tide. (After U. S. Coast and Geodetic Survey.)

to exist there. Accordingly, it seems likely that the outer shelves, as well as the inner, are products of erosion and not of deposition. Currents are probably sufficiently strong to prevent deposition on the shelves of the great quantities of the fine sediment carried out by the waves from the shoal inner portions. This fine sediment appears to be mostly swept off the shelf, and it probably lodges on the continental slopes down which much of it may slide into the abysmal depths of the ocean."

Beaches and barriers. The loose material, ranging in size from very fine to large boulders, which is shifted and ground up by the action of the waves, undertow, and shore currents, is called the *beach.* It consists of the zone of rock fragments within reach of the waves along the shore. "Its lower margin is beneath the water, a little beyond the line where the great storm waves break. Its upper margin (on shore) is at the level reached by storm waves, and is usually a few feet above still water" (Chamberlin and Salisbury). The uppor portion of the beach consists generally of coarser material, while its lower, or constantly under-water portion, is made up of finer material. Beaches are, as a rule, not prominently developed at the bases of sea cliffs, but (except along very young coasts) they are well-developed generally around the shores of coves and recesses of the coasts (Fig. 302). Where the land slopes down to the sea gently, as on coastal plains, beaches are often also finely developed.

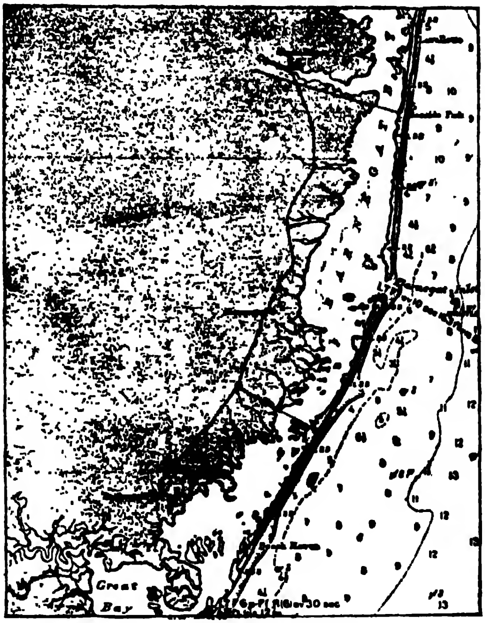

FIG. 305. Barrier beaches along the New Jersey Coast in 1916. Depths in fathoms below low tide. (After U. S. Coast and Geodetic Survey.)

Where the sea bottom slopes very gently from the shore, materials which are derived by inflow of streams and spread over the bottom by currents and undertow may be acted upon by waves which drag the bottom and break some distance from the shore. The breaking of such waves causes the water (not merely the wave form) to rush forward, stirring up and dragging along sediment. Rip currents and undertow carry back much of the material which tends to accumulate in an offshore zone where the forward rush and the reversed movement about counterbalance. A long *barrier beach* or a series of *barrier islands* thus builds up some dis-

tance out, parallel to the general coast line (Fig. 305). A barrier beach may be built up to the surface of the water by wave action and then increased in height by wind action, forming the sand into dunes. Such barrier beaches are prominently displayed along the Atlantic and Gulf Coasts of the United States from New Jersey to southern Texas.

The water of the area between the barrier and the shore is called a *lagoon* or *sound*, depending upon its size. Its water is seldom more than 10 or 20 feet deep. Lagoons are often converted partly into marshes either by accumulation of sediment from the land or by vegetation, or by both. Atlantic City, New Jersey, is built upon a barrier beach bordering such a lagoon.

Spits and bars. When a shore current, carrying sediment, comes to a cove or a narrow embayment on the coast, it tends to keep to its course rather than to follow the shore of the embayment. The sediment-laden

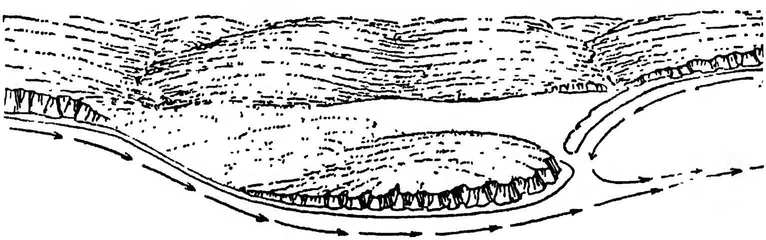

FIG. 306. A diagram to show how a bar may be built across a bay, by an eddy, in a general direction opposite to that of the main shore current. Arrows indicate directions of currents. The long bar almost completely enclosing San Diego Bay, California, excellently illustrates this principle. (After W. M. Davis.)

shore current thus moves into deeper, quieter water where its load is deposited, and thus a *spit* builds out from the shore (Fig. 303). When the current moves across the mouth of an embayment, the spit may continue to extend until it nearly or quite closes the embayment. It is then called a *bar* (Fig. 308).

Deltas. The building of deltas, often of large extent, into the sea (or into lakes) by rivers, under certain conditions, has already been considered in the discussion of stream deposition in Chapter VIII. Sea deltas are built of land-derived materials carried in by rivers, but they may be regarded, in a real sense, as marine deposits because they build out into the sea.

Wave-built terrace. Where sea cliff and wave-cut terrace are both being eroded by wave action, the loosened materials are ground up and,

in large part, may be gradually shifted over the bottom to the deeper water at the seaward edge of the wave-cut terrace and there deposited. A very considerable *wave-built terrace* may be formed by this process in the course of time (Fig. 295). The continental shelf may thus be a combination of wave-cut and wave-built terraces.

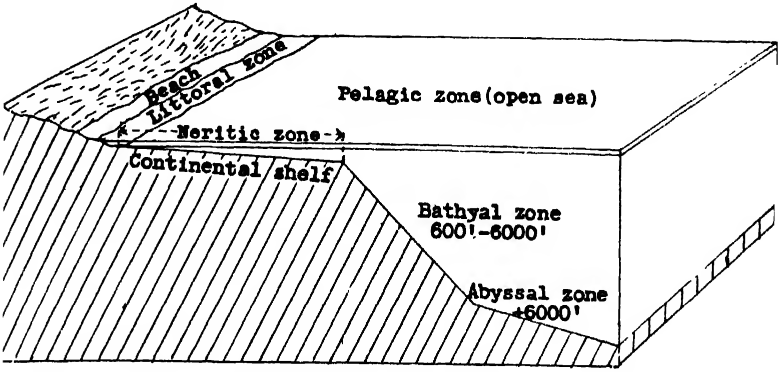

FIG. 307. Diagram showing the marine life zones. The littoral zone is a narrow area between high and low tides where, because of difficult, changing conditions, there are comparatively few organisms. The neritic zone extends from lowest tide level to the outer border of the continental shelf where the water is usually about 600 feet deep. This zone, because of good light and abundant food supply, is inhabited by organisms in great numbers and variety of forms. The bathyal zone extends from the edge of the continental shelf down to a depth of about 6000 feet. Only the upper part of this zone gets light, and so there are few plants, but there are animals in considerable variety. The abyssal zone includes the deep sea below 6000 feet in depth. Because the cold, dark water is there under such great pressure, only comparatively few highly specialized animals live in this zone. The pelagic zone, or open sea, extends shoreward as far as the littoral zone and down as far as light penetrates. This is the vast area of countless floating and swimming forms of life.

Shallow-water features, such as sea cliff, wave-cut terrace, beach, barrier, bar, and spit, are often preserved for some time after elevation of the shallow sea-bottom into land.

Deep-sea Deposits. The deposits on the deep-sea bottom, down to depths of two and three miles, are very largely organic, that is, mainly shells and other remains of organisms which have fallen to the bottom from near the surface of the sea as already explained. The most common of such deposits are the deep-sea *oozes* which are made up of the remains and shells of tiny animals and plants. Such oozes cover about 60,000,000 square miles of the deep-sea bottom.

At depths greater than two to three miles, a peculiar red clay is the primary deposit. It is very widely distributed, covering an area of 55,000,000 square miles, or an area as large as all the lands of the earth. Some remains of organisms are mixed with the clay, but since most of the shells are limy and very thin, they are dissolved without reaching the bottom in the very deep water which is not only under great pressure, but also rich in carbonic acid gas.

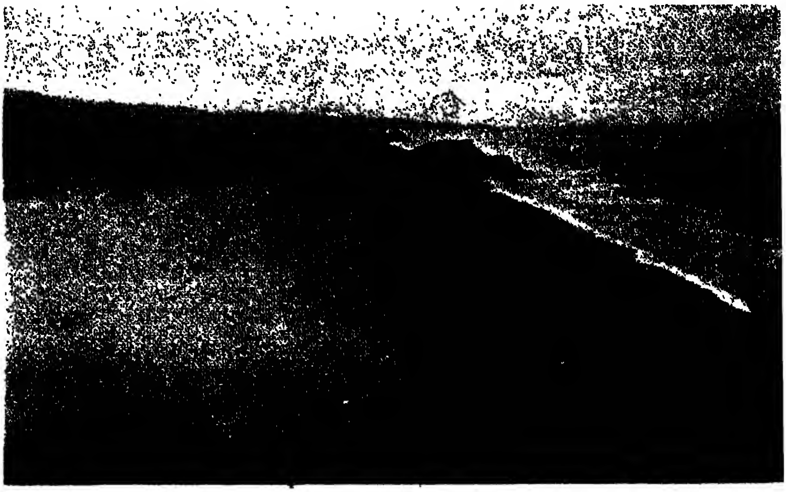

Fig. 308. A fine example of a bar built across an embayment. Coast of California, between Eureka and Crescent City. (Photo by W. G. Johnson.)

The deep-sea deposits, both oozes and red clay, do, however, contain some land-derived and other materials. Thus off the west coast of Africa some dust from the Sahara Desert is known to fall into the deep sea. Volcanic dust is often carried many miles and deposited in the deep sea, particularly in the southern Pacific Ocean. Bits of porous volcanic rock called *pumice* sometimes float long distances on the ocean before becoming sufficiently water-soaked to sink. Icebergs often drift far out from the polar regions over the sea, and, on melting, the rock débris which they carry is dropped to the sea bottom. Also particles of iron and dust from meteorites ("shooting stars") have been dredged from the deep sea.

One important geological significance of the abysmal deposits is the fact that nowhere on any continent, among the rocks of all ages as old

at least as the earliest Paleozoic, do we find any typical, deep-sea deposits. There is, therefore, no evidence that deep-sea water ever spread over any considerable part of any continent, and this is so in the face of the fact that abundant marine deposits of shallow-water origin show that shallow seas have at various times spread over large portions of continents. There has been, then, a strong tendency for the continents to maintain approximately their present positions for many millions of years.

Types of Original Shorelines

Shorelines originate in many different ways. The following very brief summary of a genetic classification of shorelines set forth by D. W. Johnson in his book, "Shore Processes and Shoreline Development," will give the reader a good idea of the main types. Changes which shorelines undergo after they originate are, of course, not included.

I. *Shorelines of Submergence* produced when the surface of the water comes against a partially submerged land area.
  - (a) *Ria shorelines* formed by partial submergence of a stream-dissected land mass with more or less drowning of the river valleys. Good examples are Chesapeake Bay and the coast of Brittany.
  - (b) *Fiord shorelines* formed by partial submergence of a region of glacial valleys or troughs like those of Norway and southern Alaska.

II. *Shorelines of Emergence* produced when the water surface comes against a partially emerged sea or lake floor. The typical case is a coastal plain shoreline like that of Texas or southern New Jersey.

III. *Neutral Shorelines* whose main features are independent of either submergence of a former land area or emergence of a former under-water surface.
  - (a) *Delta or alluvial plain shorelines* produced by the building forward of a delta or broad alluvial slope into a lake or the sea. Many examples.
  - (b) *Volcano shorelines* where an active volcano builds up its cone above a water surface. Many examples in the south Pacific Ocean.

(c) *Coral reef shorelines* where corals build upward from the sea floor or outward from a land mass. Many examples in the south Pacific Ocean.

(d) *Fault shorelines* where fault blocks are depressed enough to allow sea or lake water to come against the scarp. Fine examples near Wellington, New Zealand.

IV. *Compound Shorelines* which involve important characteristics of at least two of the preceding classes.

Normal Cycles of Shoreline Development

Cycle Inaugurated by Submergence. If a portion of a relatively rugged (mature) land surface is submerged under the sea, a very irregular, deeply indented or embayed shoreline results because of the entrance of tidewater into the valleys. Usually some islands are left opposite the headlands. Such drowned valleys are *estuaries*, or, if they are deep and narrow as a result of glacial erosion, they are *fiords*. Chesapeake Bay and Delaware Bay are good examples of estuaries, and excellent fiord coasts are those of Norway and southern Alaska. In the following discussion of stages in the history of such a shoreline, it is assumed that there is no movement of the land up or down to interrupt the normal cycle.

A newly formed coast of the kind just described (Fig. 309, A) is attacked by the sea waves which at first make it rougher and more irregular in detail. Then the islands are eroded away, the headlands are cut back somewhat, and sea cliffs are formed. This may be called the *youthful stage* (Fig. 309, B).

Next, a shallow-water shelf is cut by the waves, the headlands are cut back farther, bars are built across the embayments, and the latter begin to fill with sediments (Fig. 309, C).

Then the shoreline is cut back much farther, a very prominent sea cliff forms along most or all of the shore, and there is developed an extensive wave-cut shelf with sediment deposited beyond it perhaps in part as a wave-built terrace. The remaining sediment-filled embayments are now largely or wholly obliterated. This is the *mature stage* (Fig. 309, D). The waves then attack the whole rocky shore, and the shoreline develops irregularities or indentations because some (weaker) rock masses are cut back faster than others. Such indentations are, however, almost never comparable in size to those produced by the sinking of the land at the beginning of the cycle of subsidence.

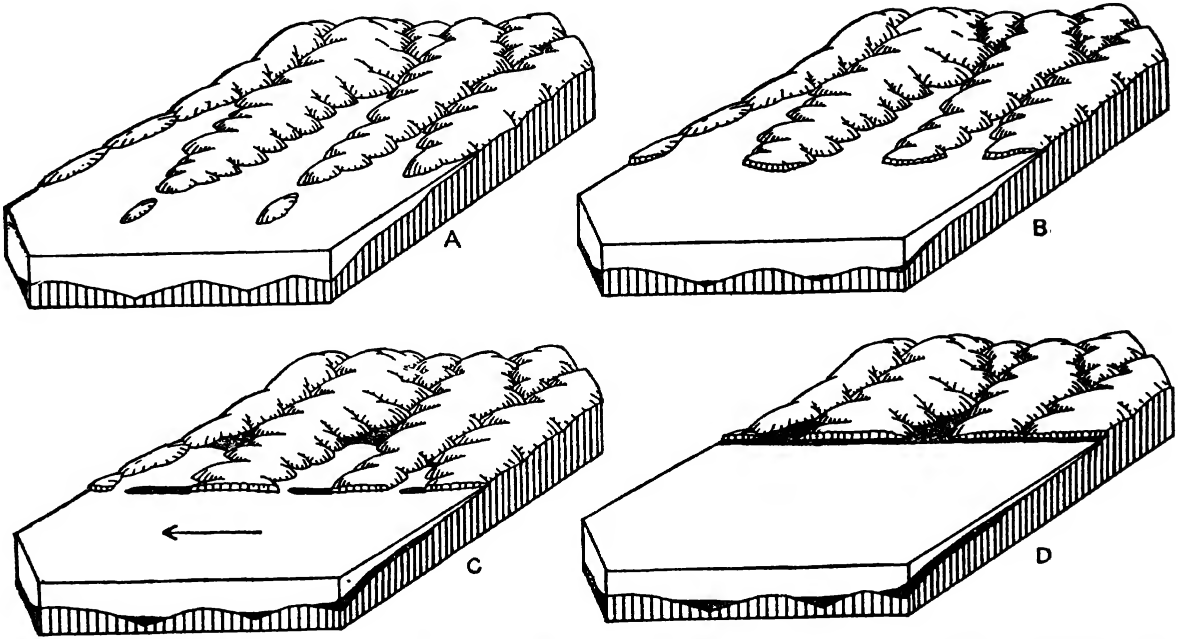

FIG. 309. Block diagrams illustrating some stages in the history of a shoreline starting with submergence of part of a mature land area. A, irregular, deeply embayed shoreline, with some islands, due to drowning of valleys; B, sea cliffs formed, headlands truncated, and islands removed; C, greater truncation of headlands, bars across the bays, and bays partly filled with land-derived sediment; and D, shoreline cut back still more, prominent and extensive sea cliffs, continuous beach, and a wide wave-cut shelf. The arrow indicates direction of shore current. (Modified after D. W. Johnson.)

In the *old-age stage* the wave-cut terrace becomes wider and wider, although at a diminishing rate because of weakened wave power on account of frictional drag where waves travel over such a wide shallow-water terrace. The sea cliff gradually becomes so faint as to scarcely merit its name any longer not only because of the weakened wave power, but also because the land is worn down to very low relief by this time. Also, because of the low land-relief, little land derived sediment comes into the sea, and so the process of clearing loose material off the wave-cut shelf is more effective than ever. There is, however, no important difference in processes involved in the mature and old-age stages.

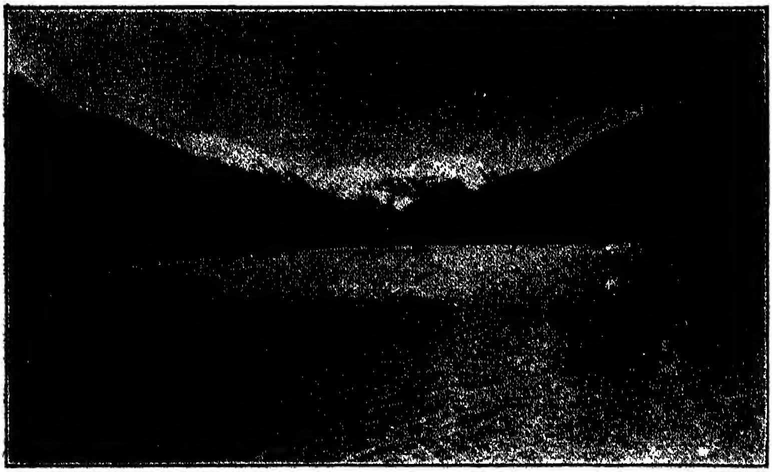

FIG. 310. A fiord. Grenville Channel, British Columbia. (Photo by W. W. Atwood for the U. S. Geological Survey.)

The *final stage* is reached when the whole land mass attacked by the sea is reduced to *base level of wave erosion,* or, in other words, when a rock platform is cut everywhere to the lower (depth) limit of wave erosion and sediments have practically all been swept off the platform. The depth of water over such a platform is usually several hundred feet, the maximum probably being about 600 feet. This final stage may be likened to the peneplain stage in the normal cycle of stream erosion.

It should be remarked that the final or even the late old-age stage involving a really extensive area is largely a theoretical consideration. This is because the widening of the wave-cut shelf gradually becomes so slow, on account of steadily weakening wave power, that, long before

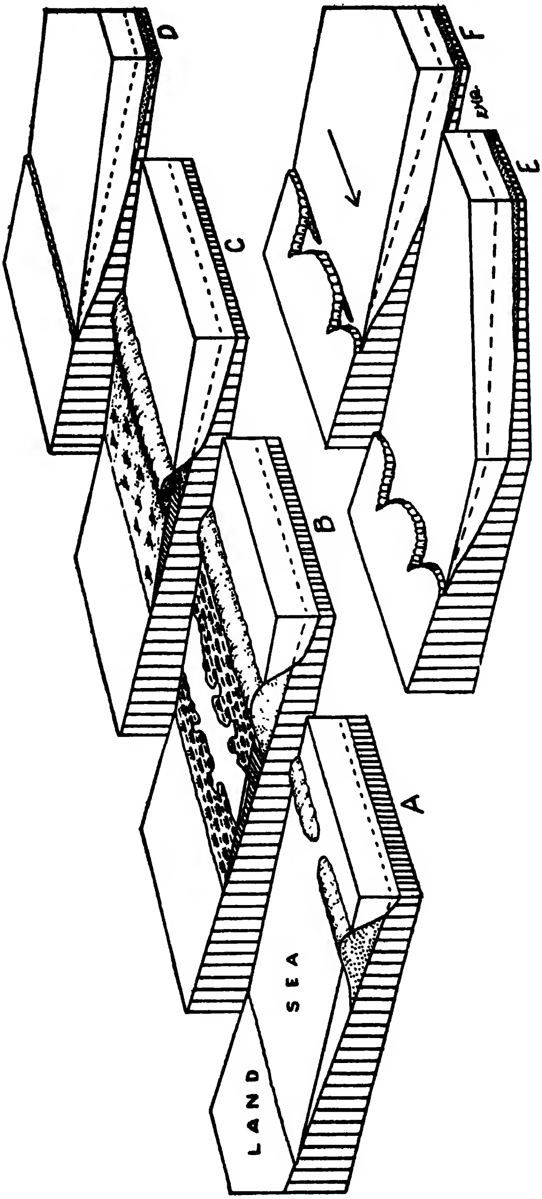

Fig. 311. Block diagrams illustrating some stages in the history of a shoreline starting with part of a flat sea bottom emerged into land. A, barrier beach (offshore bar) being formed; B, barrier beach completed with salt water and swamps between it and the land; C, barrier beach being eroded on the seaward side and migrating landward; D, barrier beach cut away by wave action and sea cliff formed; E, small bays cut out of weaker rocks; and F, bars formed across the bays. The arrow indicates direction of shore current. (After W. M. Davis, in part, with modifications.)

a land mass of even much smaller than continental size is planed away, diastrophism (particularly uplift) is almost sure to interfere with the cycle.

Partial submergence of a nearly flat land area would, of course, inaugurate a regular or relatively straight shoreline free from indentations. In such a case the shoreline cycle is essentially the same as that inaugurated by uplift of flat sea bottom into land. This cycle is described under the next heading.

Cycle Inaugurated by Emergence. If part of relatively flat sea bottom emerges into land, the resulting shoreline is, of course, regular or almost straight and free from indentations because the line of contact between the two practically flat surfaces makes it so. In the following discussion of stages in the history of such a shoreline, it is assumed that diastrophism does not interfere with the normal cycle.

If emergence produces a gently sloping coastal plain with very shallow water offshore for some distance, the waves, due to strong frictional drag on the bottom, have practically no erosive effect upon the new shoreline. Instead, the waves, by their heavy bottom drag, transport loose materials to a zone some distance offshore where a barrier beach (or offshore bar) is built up (Fig. 311, A, B). A lagoon, usually with marshes, is thus formed between the original shore and the barrier beach. The coasts of much of Texas and New Jersey are in this stage of *youth.* The depth of water on the seaward side of the barrier beach is increased enough so that waves cut into the barrier. Some of this material, especially during storms, is carried out to sea, but some of it is carried over the barrier beach. In this way the barrier migrates toward the land, and meantime the lagoon becomes partly or wholly filled with sediment and marsh-plant remains.

In time the barrier beach is completely removed, and the water offshore is then deep enough for the waves to attack the whole original shoreline. According to Johnson the shoreline is now mature (Fig. 311, D).

Indentations may then develop where some (weaker) rocks are cut back faster than others, but such irregularities are small (Fig. 311, E). Bars may be built across such indentations temporarily, but, during the *old-age stage* to the *final stage,* the history is practically like that of the same stages in the cycle inaugurated by partial submergence of a mature land area as described above.

Fig. 311, clearly illustrating the main stages of this cycle to ma-

turity, should be carefully studied in connection with the above statements.

In case emergence leaves sufficiently deep water offshore, a barrier beach does not, of course, develop, but wave cutting starts right on the new shoreline and succeeding stages are much like those of the latter part of the cycle described in the preceding paragraph.

Comparison of Coastal and River Cycles. "There are several points of similarity between coastal cycles and the erosion cycles of rivers. First a river system roughens the surface of its basin, increasing its relief; finally it reduces it to a smooth plain, near sea level. As indicated above, waves and currents normally increase to a maximum the irregularities of a coast, and finally reduce them to a minimum. An essential difference is that the irregularities of the river basin are vertical irregularities, while those of the shoreline are horizontal. In each case the cycle of development is introduced by diastrophism" (Blackwelder and Barrows).

It must, of course, be borne in mind that the normal cycles of shore development, like those of river erosion, are often interrupted either by uplift or by sinking of the land. It would involve us in too much detail, for our present purpose, to discuss the various important changes which result from such interruptions by diastrophism.

Submarine Canyons

Dr. F. P. Shepard has kindly prepared the following statement for the writer:

"It has been known for almost a century that there are submarine valleys off various coasts of the world. The surveys of the sea floor in recent years have shown that these features are very numerous and of surprisingly large dimensions. Some of these are cut as deeply into the surrounding ocean floors as are the greatest of land canyons into the surrounding land surfaces. The walls of these marine canyons are as steep as the walls of land canyons. They can be traced out for many miles from the coast and to depths of as much as 12,000 feet below sea level. Some of the most impressive of these canyons are found off the coast of California (Fig. 312).

"These valleys of the sea floor have been variously explained as the result of faulting or folding or submarine currents, but their characteristics are much more those of river-cut valleys. They resemble river

valleys in their sinuous courses, their dendritic tributary systems, and their V-shaped transverse profiles. It seems probable that they were formed at times when the coastal margins were greatly elevated (or sea level lowered), allowing streams to truncate the steep outer slopes. Subsequently submergence carried these canyons down to the great depths at which they are now found. Since submergence marine mud flows may have prevented the filling of the drowned canyons by the sediments washed out from the land."

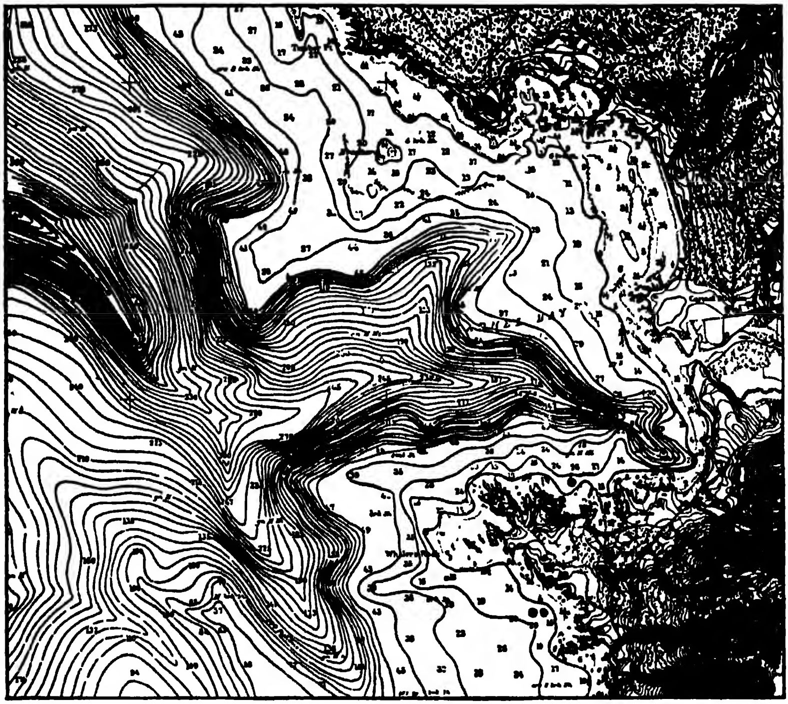

Fig. 312. The submarine canyon off the Carmel River, California. Note the dendritic pattern of the canyon. Figures show depth in fathoms. Contour interval, 10 fathoms. (After F. P. Shepard and H. H. Robinson.)

Islands in the Sea

Sea islands range in position from those very close to shore to others which lie in the midst of a great ocean. They vary in size from mere points of rock to those of very great size, for example, Australia, which may be said to have continental dimensions.

Among the most common modes of origin of sea islands are: earth-crust movements; wave erosion; wave deposition; volcanic action; and the action of organisms. Brief explanations of a few examples will serve to make the involved principles fairly clear.

Fine examples of islands formed by earth-crust movements are those off the coast of southern California. They were parts of the mainland until in very recent geological time (present period), as proved by fossil remains, etc., when they were separated by a general submergence of the region, leaving the higher seaward portions projecting as islands. Since then they have been affected by other movements, as for example San Clemente which has been elevated hundreds of feet. Many of the islands along the coast of Maine and of southern Alaska are direct results of general coastal subsidence. The great island of Australia was cut off from the Asiatic Continent about the beginning of the present (Cenozoic) era, as proved by a comparison of the fossils of the island with those of the adjacent mainland. All such islands have, of course, been attacked and modified more or less by wave action.

Many islands, generally of small size, have originated by isolation of parts of a rugged coast by wave erosion, as already outlined. Sometimes the neck of a headland or even a peninsula of considerable size may be cut through by wave action, leaving an island. Many examples occur off the coast of Maine and southern Alaska where islands and headlands resulting from submergence have been thus affected.

Relatively long, narrow islands, called barriers and barrier islands (already described), result from wave deposition offshore where the bottom is very gently inclined seaward. Numerous examples occur on the Atlantic and Gulf Coasts of the United States from New Jersey to southern Texas.

Eruptions on the sea floor may continue not only until the volcanic materials are built up to, but also far above, the surface of the sea. Wonderful examples of islands thus constructed are the Hawaiian Islands which rise out of the midst of the Pacific Ocean where the depth of the water is several miles. One of these islands is about 80 miles wide with two volcanic cones, each nearly 14,000 feet high. The total up-building by volcanic action to form this island has been, therefore, fully 30,000 feet. The Azores, most of the West Indies, the Aleutian Islands, and many of the East Indies are also islands of volcanic origin.

Oceanic islands formed by the action of organisms are mainly *coral islands*. Corals are tiny, low-order animals which build up limy skele-

tons by secretion of lime which is dissolved in sea water. Under favorable conditions, countless myriads of corals live in colonies. Extensive limestone deposits are formed by the gradual accumulation of their skeletons. Corals live mostly in clear sea water, not cooler than 68° F., and not much deeper than 150 to 200 feet. They thrive best where freely exposed to waves and currents which supply the food.

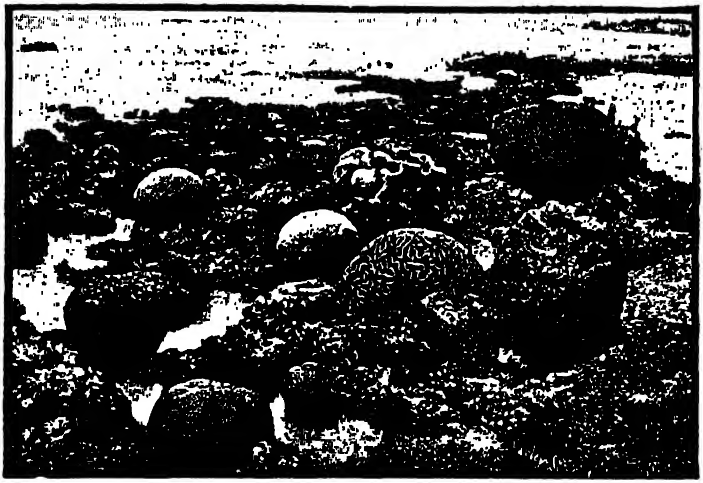

FIG. 313. Part of a great coral (barrier) reef of Australia at low tide. (After W. M. Davis.)

Chains of islands or long, narrow belts consisting of accumulated coral remains are called *coral reefs*. The greatest reefs are hundreds of miles long, as for examples on the northern side of Cuba and on the northeastern side of Australia (Fig. 313). A coral reef attached to the shore is called a *fringing reef;* one situated offshore some distance, and roughly parallel to the shore, is called a *barrier reef;* and a more or less circular reef enclosing a lagoon is called an *atoll.*

Several explanations of barrier reefs and atolls have been propounded. Thus a fringing reef attached to a roughly circular island would, by slow subsidence, accompanied by coral up-building, be transformed into a barrier reef, which on complete sinking of the island, would become an atoll. This is Darwin's theory, illustrated by Fig. 314.

According to Murray, corals growing on a submerged platform or inland truncated by the waves would build upward and outward (be-

cause of more favorable food supply) to the sea surface to form an atoll. A barrier reef might develop also parallel to a shore without subsidence. In this case a fringing reef might build steadily outward and upward from near the shore, while the portion between its outer side and the shore, not favorable to coral growth, would be scoured out and dissolved by tidal currents.

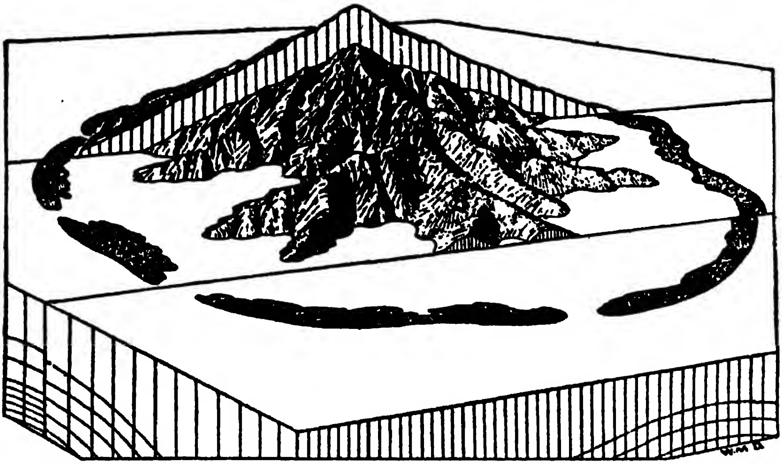

FIG. 314. A three-block diagram of a subsiding volcanic island and its upgrowing coral reef. Back block, the fringing reef; middle block, the barrier reef; and front block, the atoll. (After W. M. Davis.)

Daly believes that most coral reef foundations have remained about stationary for a long time; that only fringing reefs formed before the Ice Age; that the sea level was lowered 200 to 250 feet by withdrawal of water from the sea to form glaciers; that the water was then cold enough to retard greatly or kill most coral growth; that the waves of the lowered sea wore away the dead reefs, cut shallow-water platforms around the larger islands, and left only wave-cut platforms at the sites of small islands; and that, as the glaciers melted, sea level rose, the water grew warmer, and renewed coral activity built up reefs on the platform margins, leaving lagoons within the barrier or circular reefs. Thus he explains why most of the larger lagoons have about the same depth (about 200 to 250 feet).

Davis, after a critical study of several theories, concluded that Darwin's subsidence theory, in a modified form, best explains most phenomena of barrier reefs and atolls. He particularly emphasized the fact that so many of the islands, surrounded by barrier reefs, are em-

bayed because of valley drowning by subsidence. It may well be that each of the three theories above mentioned is more or less correct in application according to various conditions involved in different regions.

Corals can build reefs to about the average tide level only. Storm waves may break off pieces of the reef and pile them up well above sea level. Such material may then be built up into wind-blown deposits still farther above sea level, as in the case of Bermuda. Coral islands of all kinds and sizes are very common in the southern Pacific Ocean.

Origin of the Sea

Just how the sea originated is a problem by no means satisfactorily solved. Any view which we may hold regarding the origin of the ocean must be associated closely with, and largely dependent upon, our view of the origin of the earth itself.

An older doctrine, known as the "Nebular Hypothesis," regards all the material of the sun and the planets, including the earth, once to have been in the form of a vast rounded mass of highly heated, rapidly rotating gas or nebula. As this mass slowly lost heat, it contracted and left off rings, the material of each of which condensed to form a planet. Thus the earth originated, and, due to its cooling and contraction, its original, hot, heavy atmosphere, which contained all the earth's water in the form of vapor, gradually became thinner and cooler. According to this view, the waters of the sea are essentially condensed, originally highly heated gas.

A more recent doctrine has been called the "Planetesimal Hypothesis." According to it the matter of the sun and planets was at one time in the form of a great spiral-shaped swarm of masses and particles not in a gaseous condition. A few of the larger masses attracted gradually, or gathered to themselves, most of the smaller ones, and thus the planets, including the earth, were built up. As the earth increased in size, the force of gravity increased, and various gases, including water vapor, were squeezed out to form the atmosphere. When sufficient water vapor accumulated, precipitation started, and the waters of the sea began to gather. According to this view, the waters of the sea have been forced out of the earth.

Although the origin of the sea is still a problem, nevertheless we do know that the sea has been on the earth for an extremely long time. There is evidence that sea water existed in the oldest era known to the geologist, that is, at least hundreds of millions of years ago.

CHAPTER XII

SUBSURFACE WATER AND ITS WORK

Introduction

All water which occurs below the surface of the earth may be called *subsurface water,* or *underground water,* or simply *ground water.* There is good reason to believe that water (or at least its component parts) occurs intimately associated with the rocks deep down within the earth, that is, well below the zone of fracture. This is strongly suggested by the large quantities of steam given off through volcanoes and from magmas which are poured out on the earth's surface. Our present concern is not with any such very deep-seated water but rather with ground water which occurs in the zone of fracture, that is, within the outer, crustal portion of the earth.

The source of all but a very small quantity (probably not over one per cent) of subsurface water in the zone of fracture is atmospheric precipitation—rainfall and snowfall. It has been estimated that about 1500 cubic miles of water (including its frozen state—snow) fall upon the surface of the United States yearly. One-half or somewhat more of this evaporates; about one-fourth of it flows off in surface streams; and the remaining one-fourth or somewhat less works its way into the crust of the earth either by soaking into the loose materials or by entering cracks, fissures, and other openings in the bedrock. Some of the factors which favor descent of water into the earth's crust are humid climate; dense vegetation which interferes with run-off (surface flow); gentle slopes which retard run-off; and a relatively high degree of porosity and fissuring of the rock materials.

Some conception of the quantity of ground water may be gained from the statement, based upon a careful estimate, that all the water in the soils and rocks of the first 100 feet below the surface of the United States would be sufficient to form a surface layer 17 feet thick. In the sections of the country with humid climate, the amount of water in the first 100 feet would, of course, be greater than the average. It should not be assumed, however, that anything like such a proportionate amount of water in rocks continues to depths of miles or even of thou-

sands of feet. The absolute limit of depth beyond which any very appreciable amount of ground water, in the ordinary sense of that term, can exist is only about 12 to 15 miles, depending upon the hardness of the rocks. This is because the tremendous pressure of the overlying rocks makes it impossible for very appreciable openings to exist beyond such depths. Very little surface water ever reaches such extreme depths. Most of the underground water by far occurs within a few thousand feet of the surface. This conclusion is borne out by the fact that, in deep mines in various parts of the world, little or no water is usually encountered lower than a few thousand feet. Large fissures containing water are, however, sometimes found in deep mines. Some water is known to be held in the pores of the rocks beyond depths of a mile or more.

What becomes of the water which descends into the earth's crust? A large amount returns to the surface through springs and seepages; a large amount moves to the surface by capillarity in loose rock materials, and then evaporates; plants absorb much water which is drawn up into the leaves to be evaporated; a considerable amount is removed through wells; some travels underground to emerge as springs in the sea relatively near shore, as is known to be the case in the Gulf of Mexico and in the Mediterranean Sea; some enters into chemical combination with various minerals and rocks to be held there, often for ages of geologic time; and some makes its way so far down in crevices and pores of the rocks that it remains for a very long time.

Modes of Occurrence of Subsurface Water

Water in Loose Rocks and Soils near the Surface. There are three general modes of occurrence of subsurface water: (1) In loose materials relatively near the surface; (2) in porous consolidated rock layers or formations, usually well below the surface; and (3) in cracks, fissures, and other openings in hard rocks. Loose rock formations and soils are, in most humid regions, saturated with water at greater or less depths (usually less than 75 feet) below the surface. This statement is borne out by the fact that water may be obtained almost universally from wells in such regions within 25 to 75 feet of the surface. More or less moisture, of course, occurs in the materials above the zone of saturation. In arid and semi-arid regions there is often no zone of saturation in the loose, incoherent materials just below the surface or, in case it is present, it is usually farther down than in humid regions.

The porosity of many loose soils and rocks is surprisingly high. Thus 25 to 40 per cent of the volume of common sand is pore space; in loam it is usually 40 to 50 per cent. It is clear, therefore, that one-fourth to one-half of the volume of such material, when saturated, is water.

Water in Porous Rock-layers. Very considerable amounts of water occur in more or less definite layers or formations which often extend at various angles for hundreds or even some thousands of feet into the earth. Such water-bearing layers or formations are known as *aquifers.*

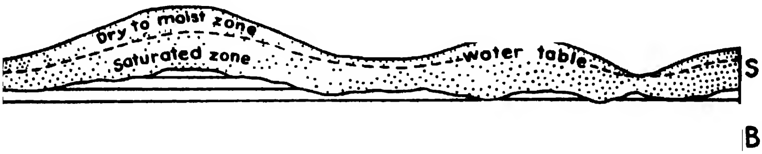

FIG. 315. Structure section showing soil (S) with water table resting upon bedrock (B).

An aquifer is usually bounded above and below by material rather impervious to water (Fig. 316). An excellent example of an aquifer on a large scale is the Dakota sandstone formation of South Dakota and Nebraska. Almost anywhere across Nebraska, a well drilled through a thick formation of clay and into the porous Dakota sandstone strikes water. In such an aquifer, water travels long distances. Thus water obtained from a well in the Dakota sandstone formation in eastern Nebraska has traveled actually hundreds of miles under the state from the eastern front of the Rocky Mountains where surface water entered the upturned and exposed edge of the porous formation.

Travel of underground water in aquifers for greater or less distances is common in many parts of the world. Such water does not of course flow freely in distinct underground channels, but rather it moves along slowly, working its way between the grains of the porous rock, and encountering much friction. The rate of motion is much slower than might be supposed. Data from various sources indicate that water in an aquifer of even coarse sandstone travels only about a mile per year. In many aquifers the rate of flow is much less.

Among the consolidated strata, sandstones and certain limestones are usually the most porous, their volumes of pore space often being 20 to 30 per cent.

Water in Cracks and Other Openings in Hard Rocks. The least amount of subsurface water occurs in the hard, bedrock formations. With the exception of the small quantity rather firmly fixed in the tiny pores of the rocks, most of such water occurs in joint cracks, fault fractures, or more or less well-defined channels. Many formations, such as granite and other types of crystalline rocks, are neither in definite layers nor porous enough to permit water really to flow through their masses. The porosity of hard, deep-seated igneous and metamorphic rocks is generally less than one per cent.

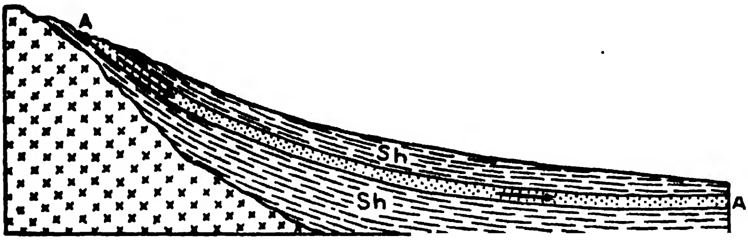

FIG. 316. Structure section showing an aquifer (A-A) lying between nearly impervious beds of shale (Sh). Arrows indicate direction of movement of water in the aquifer which may extend for many miles.

We have already learned that joint cracks are very common and usually closely spaced in all kinds of hard rocks in the outer (zone of fracture) portion of the earth's crust. Such cracks are usually more or less irregular in both direction and spacing. Fault fractures, which are not so abundant, are often rather regular and straight for considerable distances. As would be expected, the ease with which water may travel along such cracks and fractures varies greatly (Fig. 327). Many times the passageways are sufficiently long and open to permit water to follow them readily for hundreds or even thousands of feet. In the bottoms of deep canyons, water may emerge from cracks in hard rocks many hundreds of feet below where it entered the earth. It must be evident, from the above statements, not only that subterranean water does not exist in cracks and fissures in hard rocks in great amount but also that its movements are usually very irregular.

In limestone, even where it is exceptionally dense and, to a less extent, in other rocks, underground water often enlarges passageways into more or less distinct channels along which actual underground streams may flow. Such streams may reach the surface in the form of springs (Fig. 330). Echo River, which flows through the bottom of

Mammoth Cave, Kentucky, is a fine large-scale example. In a great lava region, such as the island of Hawaii, subsurface water often flows through lava tunnels, the origin of which we have already explained. In view of the facts just stated, it may be readily understood why regions immediately underlain with thick formations of limestones or lava usually have relatively few surface streams, this being because the waters easily find their way into underground channels.

The Water Table. The surface below which all openings in soils and rocks are filled with water is the *water table.* The term does not apply to a saturated layer or formation (*aquifer*) capped by an impervious layer or formation. The water table most typically lies in soil which rests upon bedrock in a humid region. In such a place surface water works downward, filling all cracks and crevices in the bedrock, and saturating the lower portion of the soil, while the upper portion of the soil is only moist. The top of the saturated zone is the water table (Fig. 315). It has already been suggested that there is no universal zone of saturation, particularly in arid regions.

The water table is very irregular, but it is generally farther under the surfaces of hills than of valleys. This is because the water at the higher levels tends to migrate, under the action of gravity, to the lower levels. After prolonged rain, the water table may coincide almost or quite with the earth's surface over a considerable area, as was the case at the time of the Dayton, Ohio flood of 1913 when the ground was nearly everywhere thoroughly soaked, causing a maximum run-off. In the soil-covered, humid portions of the United States, the water table ranges very commonly in depth from the surface to 40 or 50 feet. Springs, swamps, ponds, and lakes not infrequently mark places where the surface of the ground either intercepts, coincides with, or passes below the water table. The water table lowers steadily during long periods of dry weather, and this explains why so many wells, springs, and swamps, which are dependent upon the upper portion of the saturated zone, go dry.

Springs

Ordinary Springs.[1] The term *spring* is applied to subsurface water which issues from the ground. Springs may be divided, according to their modes of origin, into gravity and artesian springs, and, according

[1] The following statements in regard to ordinary springs are taken largely from U. S. Geological Survey Water-Supply Paper No. 255.

to the nature of the passages traversed by the water, into seepage, tubular, and fissure springs.

A *gravity spring* is one whose water is not confined between impervious beds, but flows from loose materials or open passages under the

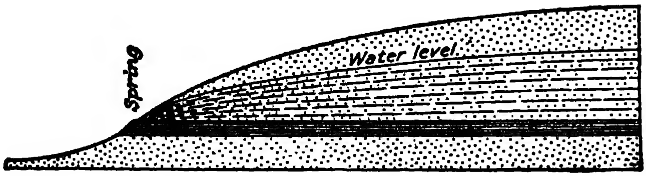

FIG. 317. Diagrammatic section illustrating a water-table (or gravity) spring. (After U. S. Geological Survey.)

action of gravity, just as a surface stream flows down its channel (Fig. 317). What may be called an *aquifer spring* is similar to a gravity spring, but its water follows a porous layer confined between impervious beds (Fig. 318).

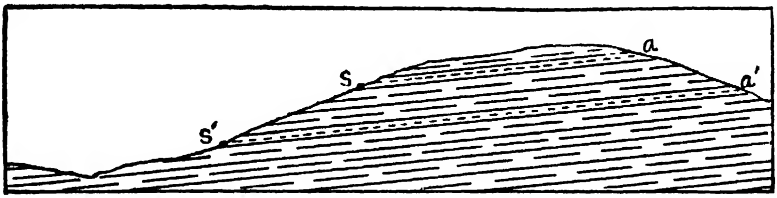

FIG. 318. Diagrammatic section illustrating aquifer springs, s and s′.

In an *artesian spring* the water is confined in impervious channels or (in porous layers) between impervious beds and is under (hydrostatic) pressure sufficient to cause it to rise to the surface (Fig. 321).

A *seepage spring* is one which the water seeps out of sand or gravel. It may emerge along the top of an underlying impervious bed, but more commonly it occurs where valleys are cut downward into the zone of saturation of a more or less uniform water-bearing deposit (Fig. 317). Seepage springs are commonly of the gravity type, but, where the channels or fissures emerge beneath beds of sand or gravel, seepages not infrequently result from true artesian springs.

Tubular springs embrace a great variety of flows, including both those in the small more or less tubular passages in (glacial) drift, and those occupying large (and small) solution channels or caverns in

soluble rocks. The channels of springs in the drift are generally established along some more or less sandy or other porous layer, or perhaps along the path left by a decaying root. In limestones and other soluble

FIG. 319. Springs issuing from a bed of gravel between layers of lava. Thousand Springs, Idaho. (Courtesy of the U. S. Reclamation Service.)

rocks, the underground passages may reach many miles in length. Some of these passages are many feet in diameter and are traversed by streams of considerable size or even by rivers (e.g., Echo River in Mammoth Cave). Tubular springs are most commonly of the gravity type, but

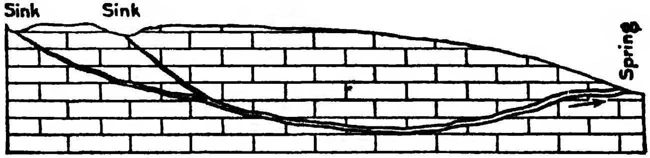

FIG. 320. Diagrammatic section illustrating a tubular spring in limestone. (After U. S. Geological Survey.)

in some cases the water may be under considerable artesian pressure in the lower parts of its channel or even at its outlet (Fig. 320).

Fissure spring is a term used rather comprehensively to include the springs issuing along bedding, joint, cleavage, or fault surfaces (Fig.

321). The distinguishing feature is a break (or network of breaks) along which the waters can pass, it being immaterial whether any considerable open space exists. Springs of this class are often distributed along straight lines for considerable distances, their position being determined by lines of fracture which are often faults. Spring water may also emerge after following a very devious course along irregular joint cracks far below the surface.

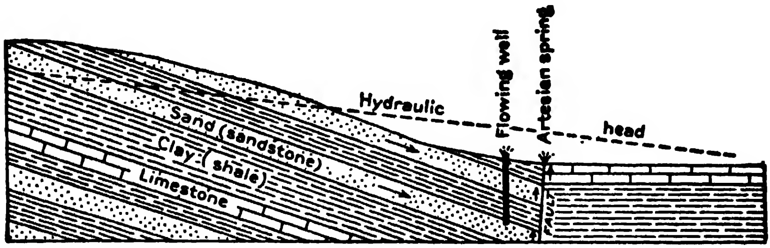

FIG. 321. Diagrammatic section illustrating a fissure (or artesian) spring. (After U. S. Geological Survey.)

Hot Springs. *Hot springs* may be regarded as those whose temperature ranges from that of the human body to the boiling point of water. The two most common causes of the heating of the waters of hot springs are the following: (1) The water may pass through masses of volcanic rocks of recent geologic age which have not yet cooled to the normal temperature of the earth's crust. Yellowstone National Park contains thousands of such hot springs (many of them boiling) where, during the present (Cenozoic) era, successive outpourings of lava covered a wide area many hundreds of feet deep. Fine examples also occur in the Lassen Peak region of northern California and in many other parts of the world. (2) Water may, where the rock structure is favorable, pass far enough below the surface to have its temperature notably raised by the general heat of the earth's interior, and then rise to the surface under (hydrostatic) pressure. It has already been stated that the temperature of the earth increases downward at the rate of about 1° F. in 50 to 75 feet. Water emerging from a depth of a few thousand feet would, therefore, be notably warm. Such springs are, however, usually not very hot, and rarely, if ever, actually boiling. They emerge usually from prominent fault fractures which extend to great depths, generally where the rocks are also much folded. There are many examples in the southern half of the Appalachian Mountains, as at Hot Springs, Virginia. Among many other examples

are Hot Springs, Arkansas; near Ogden, Utah; and in parts of southern California.

Other sources of heat of underground water may be chemical action; friction due to rubbing of rock masses against each other, as during faulting; and possibly radio-activity, but these are probably much less important than the two sources above explained.

Geysers are periodically eruptive hot springs found only in a few of the recent volcanic regions of the world, such as Yellowstone Park, Iceland, and New Zealand. They are exhibited most wonderfully in Yellowstone Park where many of them erupt columns of hot water to

FIG. 322. A group of hot springs and geysers. Yellowstone National Park. (Photo by Hillers, U. S. Geological Survey.)

heights of 25 to 250 feet at intervals varying from an hour or less to many days (Fig. 322). Old Faithful Geyser erupts once about every 70 minutes, each time sending over a million gallons of hot water, in the form of a column several feet in diameter, to heights of 150 feet or more. In 1938 it erupted to a height of 223 feet.

The general explanation of geyser action may be stated briefly as follows: The very irregular geyser tube extends downward nearly vertically into a mass of hot lava. The tube is filled with water from openings in the immediately surrounding rocks. The boiling point of the water toward the bottom of the tube is considerably greater than it is at the surface because of the pressure of the column of water. Finally, however, the hot lava causes the water to boil far down in the tube, in spite of the pressure. The first steam to form causes the whole

column of water to lift slightly, thus relieving the pressure on the superheated water far down, and resulting in a quick development of much steam which violently forces most of the water out of the geyser tube.

Mineral Springs. Water begins to take mineral matter into solution as soon as it enters the earth. Where surface drainage enters the earth,

Fig. 323. Detail view of a part of Mammoth Hot Springs. Yellowstone National Park. (Photo by Jackson, U. S. Geological Survey.)

some mineral matter is already in solution. The amount of material dissolved depends upon various conditions such as the distance the water travels, the kind of rock traversed, the pressure, the temperature, and the content of gas. In many cases, especially where the water goes but a short distance below the surface in difficultly soluble materials, the amount of mineral matter in solution may be so slight as to be unnoticed by ordinary observation. In many other cases, however, particularly where the material is relatively soluble, or where the water travels far down, much material may be taken into solution, causing the water to become more or less highly mineralized. Such water, emerging at the earth's surface, forms a *mineral spring* which may be cold or hot. Mineral springs and also wells often yield so-called hard water which contains much calcite, dolomite, gypsum, or certain other mineral salts in solution. Wells and springs in limestone regions characteristically yield hard water. Soft water usually emerges from openings

in igneous rocks, such as granite and lava, and from other rocks which contain very little limy matter. Mineral waters may contain sulphuretted hydrogen, carbonic acid gas, and other gases. A carbonated spring is highly charged with carbonic acid gas which escapes from the emerging water on account of the relief of pressure. Mineral springs may be either hot or cold. Mineral waters are often more or less medicinal in their effects.

Wells

Kinds and Depths of Wells. Water wells are sunk in various ways, the most common of which will be mentioned. Most wells by far are simply *dug* down in loose materials to a little below the water table. The depth seldom exceeds 50 feet in humid regions. Wells are often *bored* in loose materials with large augers, rotated by a power-developing machine, to depths of 100 feet, or somewhat more. Wells may also be *driven* in loose materials by forcing small metal tubes with perforated points to depths of 50 to 200 feet.

In hard bedrock formations, wells for water and oil are usually *drilled* by the percussion of a long, heavy steel weight which is raised and suddenly lowered repeatedly from a derrick or by the more recent rotary method. Many wells have been drilled to depths of thousands of feet. Many years ago a well in West Virginia showed a temperature of 172° F. at a depth of 7000 feet. In 1938 a well drilled to a depth of 15,004 feet in California showed a bottom temperature of more than 300° F. Still more recently (1947) wells have been drilled to depths of 16,668 feet and 17,823 feet, respectively, in California and Texas.

The drilling of deep wells, where samples of the rocks from different levels have been saved, has been an important aid to the geologist in rendering more precise a knowledge of the kinds, thicknesses, and structural relations of the rocks underground.

Artesian Wells. When a well, sunk to a porous water-bearing layer or formation or a crack or fissure filled with water, encounters water under enough (hydrostatic) pressure to cause it to rise more or less in the hole it is an *artesian well.* The water is often under a tremendous

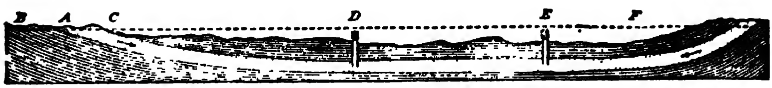

FIG. 324. Structure section illustrating flowing wells in a synclinal basin. (After U. S. Geological Survey.)

so-called "pressure head," and it may or may not flow out upon the earth's surface like a fountain.

Requisite conditions for the most common type of artesian well are the following: a porous layer between water-tight layers; exposure of at least an edge of the porous layer so that water may enter it; inclination of the water-bearing layer (aquifer) so that the water will move

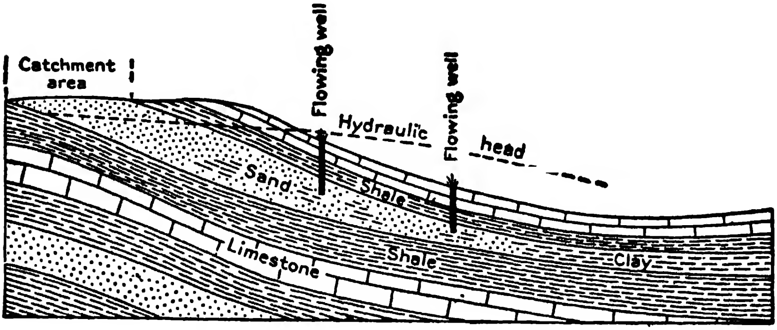

FIG. 325. Structure section illustrating flowing artesian wells in a monocline. (After U. S. Geological Survey.)

downward in it under the action of gravity; absence of a ready escape of the water at a lower level than that of the well; and an adequate rainfall to furnish the supply of water.

An aquifer like that just described may extend under a valley, and outcrop on the hills on each side as shown by Fig. 324; or it may be tilted in one direction and thin out or grade into impervious material, as shown by Fig. 325. In either case a flowing artesian well would be obtained by sinking a well through the upper water-tight layer into the aquifer. Some artesian basins are very extensive, and the water emerging from a well in such a basin may have traveled hundreds of miles underground.

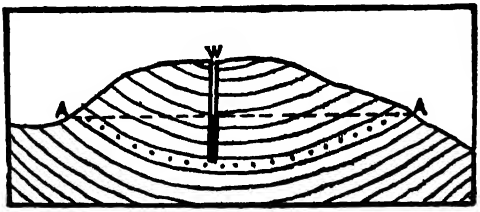

FIG. 326. Structure section illustrating a nonflowing well in a synclinal area. Dotted formation is the aquifer.

If an aquifer, lying between water-tight beds, curves downward (synclinally) under a ridge, as shown by Fig. 326, a well sunk to the aquifer from high up on the

hill would be nonflowing, although the water might rise under great pressure to a considerable height in the hole. In none of the cases described will the water rise to the level of its source (or intake) because friction during the passage of the water through the porous rock layer reduces notably the pressure, the more so as the distance increases.

Much less commonly than the case just mentioned, both flowing and nonflowing artesian wells may result where water is encountered under pressure in cracks, fissures, or channels in dense or hard rocks as suggested by Fig. 327.

Wells and Sanitation. A large percentage of the people of the United States depend upon wells for their water supply. Most of the people by far in the upper Mississippi Valley region use well water. The location of wells with reference to sanitary conditions is, therefore, of very great importance. Failure to give reasonable attention to simple, fundamental precautions is a reason for a large amount of sickness which could be avoided, especially in country districts.

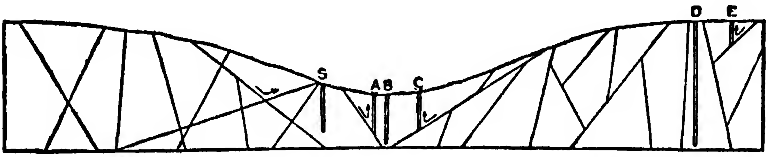

FIG. 327. Diagrammatic section showing springs and flowing wells in jointed rocks. (After U. S. Geological Survey.)

Germ-laden water may travel surprisingly far underground. Water contaminated by barnyards, cesspools, and outhouses spreads notably on sinking to the water table in loose materials, often causing water in shallow (dug) wells close to houses and barns to become more or less germ-laden (Fig. 328). The safe well must be situated out of range

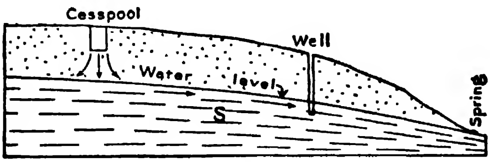

FIG. 328. Structure section showing one way by which wells and springs may be polluted. (After U. S. Geological Survey.)

of such contamination. Germ-laden surface water may also travel underground through fissures, cracks, or channels in bedrock, and con-

taminate wells and springs. Less often the surface and near-surface drainage may be down a hillside, while contaminated water may flow in the opposite direction underground in a porous layer of tilted bedrock. Even after a well has been carefully located in the light of the principles suggested, sanitary analyses of the water should be made once or twice a year to insure reasonable safety.

Work of Solution by Subsurface Water

Solvent Action of Subsurface Water. Mention has already been made of the fact that water begins to take more or less mineral matter into solution as soon as it enters the earth. Pure water has some power to dissolve mineral matter, but the carbonic acid gas and other gases and acids which it takes up from the air and from the decomposing organic matter in the soil greatly increase its solvent power. If it moves downward far enough, it also becomes warm (or hot) and gets under pressure, thus making it a more powerful solvent. The most common rock which is readily soluble in such water is limestone, which consists largely or wholly of either calcite or dolomite. Gypsum and salt-bearing strata are readily attacked. Many other minerals, even as resistant as feldspar, also are taken at least partly into solution.

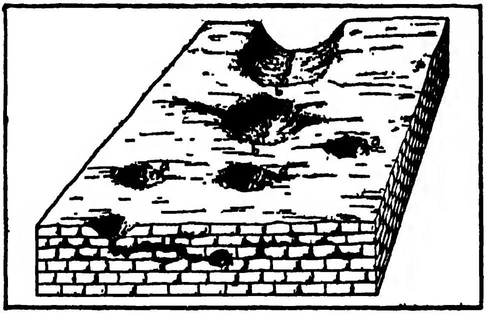

Fig. 329. Diagram illustrating the origin of caves (in black), sink holes (a), and natural bridges (b), in limestone. (After H. F. Cleland.)

Amount of Material Dissolved. One of two principal things may happen to mineral matter taken into solution by subsurface water. It is either carried deeper down into the zone of fracture and there deposited in openings of the rocks or it is brought to the surface, mainly through springs, to be carried to the sea by surface streams. The deposition of materials at lower levels is discussed beyond in this chapter. Even

roughly approximate figures in regard to the amount of material deposited at lower levels in the zone of fracture cannot be given, but the quantity is certainly large.

In the discussion of the rate of erosion of the United States, it is stated that, according to good estimates, the rivers of the country carry about 270,000,000 tons of dissolved mineral matter to the sea each year. Most of this enormous amount of material is taken into solution by underground waters and fed into the streams through springs. Underground water, therefore, contributes notably to the general process of wearing down (erosion) of the lands, because the surface streams, as they cut down, have less rock material to remove.

FIG. 330. A large volume of water issuing from an underground passage in limestone. Big Spring, Missouri.

Caves, Sink Holes, and Natural Bridges Formed by Solution. An important result of the solvent action of subsurface water, particularly in limestone regions, is the development of *caves* (or *caverns*). One of the most remarkable examples is Mammoth Cave, Kentucky. This wonderful work of nature is entirely the result of the action of underground water which has dissolved (and to some extent corraded) and carried away tremendous quantities of limestone. There are scores of miles of intricate passageways and galleries, some of them very large. A stream, called Echo River, aided by its tributaries is still carrying on the work of solution, and so the cave is being enlarged. Among other famous, large caverns similarly found in limestone are the recently discovered Carlsbad Cave, New Mexico; Oregon Caves, Oregon; Wind Cave, South Dakota; and Luray Cave, Virginia.

An opening which connects a cave with the surface is called a *sink hole.* Sink holes may be formed either by the solvent action of surface

FIG. 331. A stream entering an underground passage in limestone. Nakimu Caves, British Columbia.

water which finds its way into a cave, or by the collapse of part of the roof of a cave.

A *natural bridge* may be formed by the collapse of all but one portion of the roof of a cave. A famous example is the Natural Bridge of Virginia. Natural bridges are formed also in other ways, one of which is described in Chapter VIII.

Deposition by Subsurface Water

Cave Deposits. When water containing carbonic acid gas passes downward through a limy formation it becomes more or less lime-charged. A drop of such lime-charged water, on reaching the roof of a cave, evaporates somewhat, and gives up some of its gas, with the result that part of the lime is deposited. After hanging for a time on the ceiling, the drop of water falls to the floor where much or all of the remaining lime is deposited. Many repetitions of this process cause a long, slender, icicle-shaped incrustation of carbonate of lime, called a *stalactite,* to be built vertically downward from the roof of the cave, and a similar, though usually thicker, mass, called a *stalagmite,* to be built vertically upward from the floor. Many stalactites and stalagmites may form in a single cave, and some of them may join to form columns or pillars (Fig. 332). Wonderful and fantastic effects are thus often produced, dependent upon the manner in which the lime-charged waters trickle and spatter, as for example in the Luray Cave of Virginia; parts of Mammoth Cave, Kentucky; and Wyandotte Cave, Indiana. Stalactites and stalagmites occur in great profusion, and of great size—5 to 25 feet in diameter, and 25 to 50 feet long—in the very recently explored Carlsbad Cave of New Mexico.

Under more exceptional conditions, stalactites and stalagmites may be formed of other minerals, such as chalcedony, limonite, etc. These

are rarer and usually much smaller than those of lime because the materials are more difficultly soluble.

Fig. 332. Stalactites, stalagmites, and pillars in a cave. Oregon Caves, Oregon. (Photo by courtesy of the U. S. Forest Service.)

Spring Deposits. When underground water, highly charged with mineral matter, reaches the surface as a spring, there is a strong tendency for it to deposit at least part of its mineral load. Reduction of pressure, lowering of temperature, and escape of carbonic acid gas are among the principal factors which cause such deposition by springs (Fig. 323). Deposits of carbonate of lime are not uncommonly found around springs of even relatively cool water, where the mineral-charge is heavy. *Travertine* is a general name applied to limy spring deposits, whereas the more porous or stringy limy masses are called *calcareous tufa* (Fig. 13).

Large, hot springs are especially likely to yield extensive deposits in their immediate vicinities, an excellent case in point being the great accumulations of travertine around the Mammoth Hot Springs of Yellowstone Park (Fig. 323). The alkaline waters of the hot springs and geysers of the Yellowstone geyser basins bring much so-called *geyserite* to the surface where it accumulates. This porous material is

the same in composition as the mineral quartz. Other mineral substances are less often deposited by springs.

Deposits in the Belt of Cementation. Underground water dissolves much material in the upper portion of the zone of fracture. As the water moves downward, it becomes richer in mineral matter, and more sluggish. Ascending hot water, charged with dissolved material, loses pressure and cools. The tendency is for dissolved substances to be deposited, under such conditions, filling cracks, fissures, and openings of all kinds, even exceedingly small ones. That portion of the zone of fracture, in which most deposition of dissolved minerals takes place, is called the *belt of cementation*. Many sedimentary rocks are consolidated by cementation in this belt. Cracks and fissures filled with mineral matter from underground water solutions are called *veins* (Fig. 333). Among the very common vein-forming minerals are quartz, calcite, fluorite, and barite. Two or more minerals may occur in one vein. Where underground openings are filled only partly with mineral matter, beautiful crystals often occur.

FIG. 333. Veins of calcite in a pebble of schist.

Valuable ores, such as those of gold, silver, copper, lead, and zinc, usually have been deposited from underground water solutions and concentrated in veins in many regions. Deposition also often results where underground waters with certain substances in solution travel through various rocks or encounter solutions of other substances, thus bringing about chemical reactions which may develop insoluble substances, with resultant deposition of the latter.

Mineral-charged subsurface water may also bring about *petrification*, that is, the replacement, particle by particle, or cell by cell, of a buried shell, log, or other remains of an organism by the mineral matter from

an underground solution. In this manner the so-called Petrified Forests of Arizona (Fig. 334), of Yellowstone Park, and of other regions were formed, the petrifying material having been the very common substance called "silica," which is the same in composition as the mineral quartz.

FIG. 334. Petrified tree-trunks uncovered by erosion. Petrified Forest National Monument, Arizona. (Photo by G. P. Merrill, U. S. National Museum.)

CHAPTER XIII

MOUNTAINS, PLATEAUS, AND PLAINS

Principal Relief Features of the Land

Introduction. The earth's relief features of first magnitude are the *continents* and the *ocean basins*. The principal relief features of the continents are *mountains, plateaus*, and *plains*, known as second-order features. Many references to these have been made in the preceding pages, but our present purpose is to consider them briefly as units, with particular emphasis upon their origin and history. It has been well to reserve such a discussion of mountains, plateaus, and plains until late in our study of physical geology because it involves a knowledge of so many facts and principles of the science, more especially of diastrophism, vulcanism, and erosion, and also a knowledge of common rocks. Ocean basins have already been considered in Chapter XI.

Mountains constitute the most conspicuous relief features of the earth. The expression "everlasting hills" may seem appropriate to the layman who is impressed by the grandeur and massiveness of mountains. To the geologist, however, mountains, even the grandest of them, are known to be but transitory forms. A mountain, like an organism, has a life history which may be relatively short and simple or long and complex. Many of the most profound lessons of geology have been learned from the study of the tilted, folded, faulted, and deeply eroded rocks of the earth's crust where they are exhibited so wonderfully in mountains.

Mountains. In the commonly accepted sense of the term, a mountain is any notably elevated portion of a region. As more precisely defined, "*mountains* are conspicuously high lands which have but slight summit areas" (R. D. Salisbury). Mountains are conspicuously high in a relative sense only, that is, they stand out boldly above their surroundings. Low mountains are often called *hills*, but the distinction between these two terms is often a relative matter, usually depending upon the region in which the elevations occur. Thus in a region of low relief like Iowa, elevations of only 100 to 300 feet are sometimes referred to as mountains (e.g., Mount Vernon, Iowa), while in other regions elevations of 1000 to 3000 feet may be called hills (e.g., Berk-

shire Hills, Massachusetts). As a rule, however, elevated masses lower than a few hundred feet are not called mountains, and those higher than about 1000 feet are not called hills.

Plateaus. Tracts of relatively high land with considerable summit or near-summit areas are called *plateaus*. They nearly always rise distinctly and rather abruptly above the surrounding country on at least one side. True plateaus rarely, if ever, merge into lowlands (plains). Plateaus are usually higher than plains, but they may be considerably lower, as for example the Piedmont Plateau of the eastern United States which is much lower than the Great Plains lying just east of the Rocky Mountains. Plateau surfaces are usually more or less trenched by valleys, or even great canyons; and mountains rise above the general level of some of them.

An excellent large-scale example of a high-level plateau with a conspicuous descent on one side is the great Colorado Plateau of the southwestern United States (Fig. 4). It lies from 5000 to 11,000 feet above sea level, with a gradual increase in altitude from south to north, and it is separated from the Great Basin on the west by a steep slope (fault scarp) 1000 to 3000 feet high. It is trenched deeply by the Grand Canyon of Arizona.

Some plateaus lie between plains and mountains. Examples are the Cumberland (or Allegheny) Plateau which lies between the Appalachian Mountains and the Interior Lowland or Plain, and the Piedmont Plateau which lies between the Appalachian Mountains and the Atlantic Coastal Plain (Fig. 4).

Some plateaus, like Greenland and the Iberian Peninsula (Spain and Portugal), rise abruptly either from the sea or from narrow coastal plains, on one or several sides.

The remarkably lofty plateau of Tibet (altitude, fully 15,000 feet) is almost surrounded by mountains, as is also the plateau of central Mexico.

Plains. Tracts of relatively low, level lands are called *plains*. In actual usage the terms "plains" and "plateaus" are often confused, and, as a matter of fact, a very clear distinction between them is difficult to make. Plains are, as a rule, lower than plateaus, but there are striking exceptions, as for example the Great Plains of the United States lying at altitudes of from 2000 to 6000 feet, and gradually descending eastward into the Interior Lowland or Plain (Fig. 4). Relation to the surrounding country is a more important criterion than altitude for

distinguishing between plateaus and plains. Thus if the region known as the Great Plains were separated abruptly from the Interior Lowland, or if it were almost surrounded by mountains, some such term as "Great Plateau" would be more appropriate.

Much of the continental areas are occupied by plains. Not only are plains the simplest of land forms, but also they are the most widespread. Most of the people of the world by far live upon plains. In the United States, plains are excellently and extensively illustrated by the Interior Lowland or Plain, the Great Plains, and the Atlantic and Gulf Coastal Plains (Fig. 4). The Great Plains are remarkably smooth, but the others mentioned are considerably trenched by stream-cut valleys.

Form and Arrangement of Mountains

A *mountain peak* is a more or less cone-shaped mountain mass, as for examples Lassen Peak, Pikes Peak, Mount Rainier, and Mount Washington.

A *mountain ridge* is a relatively long, narrow mountain mass, such as the Blue Ridge and many others, often locally called mountains or ranges, in the Appalachian district.

Peaks or ridges, or both, may be grouped irregularly, as in the Adirondack and Catskill Mountains of New York. A single, large ridge may be surmounted by a number of peaks, as for example the Cascade Range. A single, large ridge may be without very conspicuous peaks at its crest, as for example the Sierra Nevada Range. Many nearly parallel ridges may be grouped into long, relatively narrow belts, as in the case of the Appalachian Range.

A *mountain range* may, from the geological standpoint, best be regarded as a single mountain ridge or group of ridges and peaks, often with more or less parallel arrangement, the material of which was built up into mountain form by a geological process (or set of processes) during a particular portion of geological time. In dealing with mountain origin and structure, the range is, therefore, a geological unit. The Appalachian Range, the Sierra Nevada Range, the Wasatch Range, the Coast Range, the Pyrenees, the Alps, and the Himalayas are good examples, although it should be borne in mind that most of these were rejuvenated after their original uplift.

A *mountain system* "consists of two or more mountain ranges, of the same (or nearly the same) period of origin, belonging to a common

region of elevation, and generally either parallel or in consecutive lines" (J. D. Dana). Thus the Laramide system includes a series of ranges in the Rocky Mountains.

A ***mountain chain*** consists of two or more systems or ranges formed at distinctly different geological times in a definite part of a continent and usually more or less parallel. Thus the Appalachian Chain comprises the whole mountain region on the Atlantic side of North America, including the Acadian Range of Nova Scotia and New Brunswick, the mountains of eastern New England, the Green Mountains, the Berkshire Hills, the Highlands of the Hudson, and the Appalachian Range.

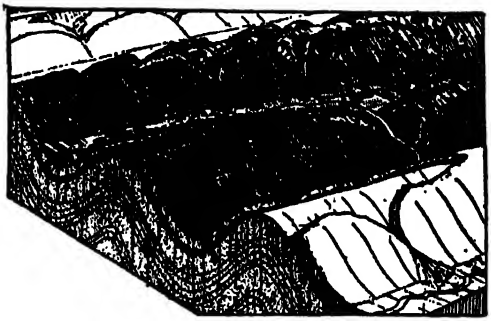

FIG. 335. Diagram and section showing slightly eroded mountain folds. Jura Mountains, Switzerland. (After W. M. Davis.)

A *cordillera* is a grand combination of chains, systems, and ranges in one general portion of a continent. The North American Cordillera includes all the mountains from the eastern face of the Rocky Mountains to the Pacific Ocean.

Origin of Mountains

Folded Mountains. *Character, origin, and structure of the materials.* Most of the great mountain ranges of the earth belong in the category of so-called *folded mountains.* Folding of strata, accompanied by general uplift and followed by erosion, is the most important of the various modes of origin of mountains. A good idea of the general character, origin, and structure of the materials of a typical folded range may be gained from the consideration of a carefully studied example, such as the Appalachian Range.

Even a casual trip across the Appalachian Range would reveal the fact that the rock materials consist very largely of common kinds of stratified rock, that is, sandstones, conglomerates, shales, and limestones. It would also be evident that the thickness of the strata must be measured by thousands of feet. As a matter of fact, careful determinations

have shown that the strata of the Appalachians were deposited originally under water, layer upon layer, to a maximum thickness of 25,000 to 35,000 feet. The tremendous thickness of such a pile of strata clearly leads to the conclusion that the deposition must have continued for

FIG. 336. Structure section showing deeply eroded folded strata in the Appalachian Mountains of West Virginia. C = Cambrian; O = Ordovician; S = Silurian; D = Devonian. Length of section, 18 miles. (After Darton, U. S. Geological Survey.)

millions of years. Not only are the rocks of the Appalachians water-laid sediments of great thickness, but also they were deposited mostly under sea water as proved by the fact that they contain numerous fossil remains of typical marine animals.

The strata of a typical folded range, like the Appalachians, are largely or wholly of shallow-water origin, that is, they were laid down

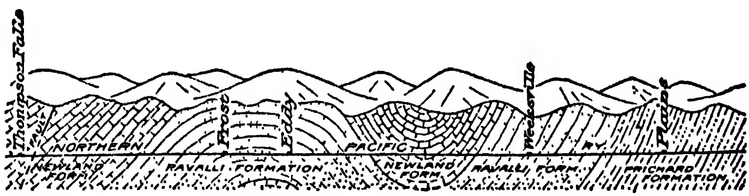

FIG. 337. Structure section 30 miles long showing deeply eroded folds. Rocky Mountains of Montana. (After U. S. Geological Survey.)

on the floor of a relatively shallow sea. This is proved by the very nature of the materials, particularly the sandstones and conglomerates; by the types of animals represented in fossil form; and by certain markings on many strata, such as ripple marks, mud cracks, etc. Since the strata are of shallow-water origin, and since they are piled up to a great thickness (many thousands of feet), it is obvious that the sea floor upon which the sediments accumulated must have subsided during the process of deposition. Such deposition of sediments usually takes place in a great down-warp, or subsiding trough, generally hundreds of miles long and 75 miles or more wide, known as a *geosyncline* to distinguish it from an ordinary syncline.

The folded strata of a typical folded range are nearly always arranged in relatively long, narrow belts or zones. This is because

they consist very largely of land-derived materials which were deposited in shallow water along the margin of a land area. This is in harmony with the well-known fact that, at the present time, land-derived sediments (gravel, sand, and mud) are deposited almost entirely within 100 to 200 miles of the continents. We must, therefore, think of the

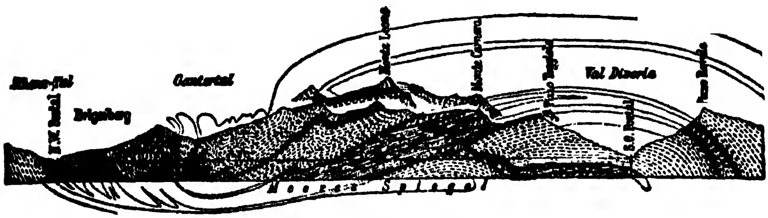

FIG. 338. Structure section showing the deeply eroded, very highly folded Alps along the line of the Simplon Tunnel. Length of section, 16 miles. (Kayser, after Heim.)

site of a typical folded range as once having been a subsiding, marginal sea bottom upon which sediments piled up layer upon layer for millions of years to a thickness of many thousands of feet.

One of the most strikingly evident features of a typical, folded range is that the strata are not in essentially horizontal position as they were when they were deposited, but that they have been much disturbed and thrown into folds. Single folds range in length and width from less than a few feet to miles (Figs. 73 and 337). The degree of folding varies from gentle anticlinal and synclinal structures to overturned and recumbent folds and even to compressed isoclinal folds (Figs. 66 and 71). Such folded structures were, as explained in Chapter V, developed by a tremendous force of lateral compression within the zone of flowage of the earth's crust. The folds are now exposed as a result of removal of overlying material by subsequent erosion. The main axes of the folds, with some minor exceptions, extend essentially parallel to the main trend of a folded range. This is because the force of compression was exerted at right angles to the trend of the range.

A high degree of folding of strata results in a considerable amount of earth-crust shortening. This is because the belts of once horizontal strata are crumpled into much narrower zones. It has been estimated that the crustal shortening caused by the Appalachian folding across southern Pennsylvania was fully 26 miles. In the very severely folded Alps the shortening is much greater.

Brief history of a folded range. In dealing with a cycle of erosion or topographic development, we used the terms infancy, youth, maturity, and old age. In a somewhat similar manner we may use a biological analogy in dealing with the evolution of a typical, folded range. First there is the *embryonic stage* during which the sediments accumulate upon the marginal sea floor. This stage is usually very long—millions of years at least.

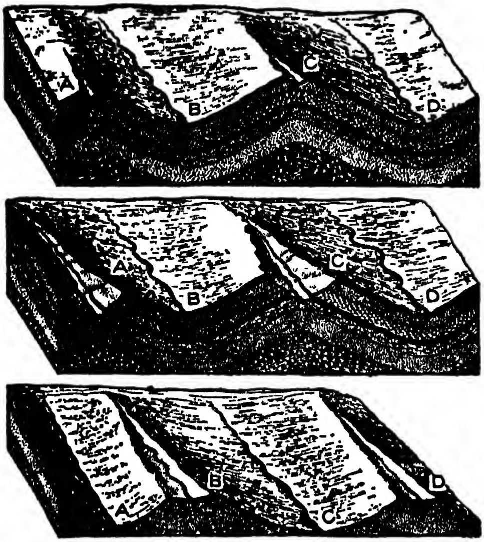

FIG. 339. Block diagrams showing how erosion may cause anticlinal valleys and synclinal mountains. (From Tarr's *New Physical Geography* by permission of the Macmillan Company.)

Next comes the *birth* of the range when the strata are subjected to lateral pressure, somewhat folded, and raised partly out of the sea.

During the *youthful stage* the mountain range grows, that is, it increases in altitude, and the folding becomes more complex because the compressive force is still very active. The increase in height takes place because the constructive force of uplift is greater than the destructive force of erosion, which latter already operates to cut down the range.

The *mature stage* is reached when the upbuilding process is about equaled by the tearing-down (erosive) process. It is during this stage that the range exhibits its greatest altitude and its maximum ruggedness of relief.

During the *old-age stage,* the upbuilding process either greatly diminishes or ceases altogether, and the tearing-down process of erosion causes a steady reduction in the height of the range.

Finally the *extinction* of the range, as a conspicuous relief feature, is reached when erosion has reduced it to the condition of a peneplain.

It is evident, from the above statements, that, even before a range attains its maximum height above sea level. it is considerably modified

by erosion. When the first fold appears above the water, erosion begins its work and continues with increasing vigor as the mountain mass gets higher. Thus there is a great warfare between two great natural processes—building up and tearing down. As long as the building-up process predominates, the mountain range increases in height. When the tearing-down process predominates, the height diminishes. The mountain range is obliterated after uplift ceases.

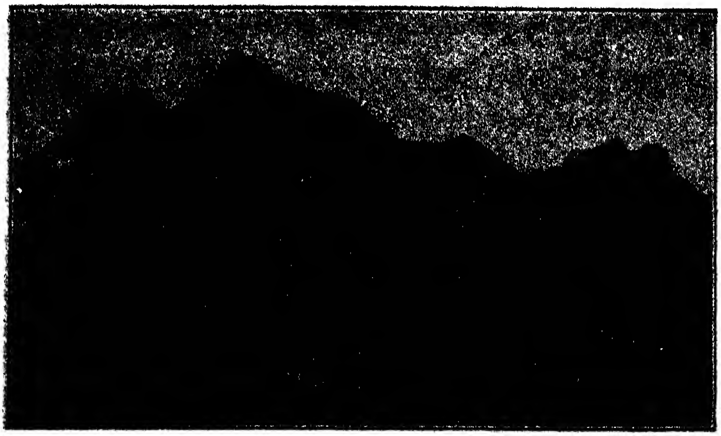

FIG. 340. A mountain ridge carved out of highly inclined, folded strata. Height of mountain, about 2000 feet. Heaven's Fold, Montana. (Photo by Chapman for U. S. Geological Survey.)

We have here a fine illustration of one of the remarkable procedures of nature. After millions of years of work, resulting in deposition of strata, layer upon layer, to a thickness of many thousands of feet, a tremendous compressive force is brought to bear, the strata are folded, and a mountain range is literally born out of the sea. The sediments derived from the erosion of the range are carried into the nearest sea to accumulate again, and after long ages they may be folded up into another mountain range. Thus much of the sediment in the folded strata of the Coast Range Mountains of California was derived from the wearing down of the Sierra Nevada Range, and this material is now being carried into the Pacific Ocean. We learn from this that the many mountain ranges of the earth are by no means all of the same age. To illustrate, the Appalachian Mountains are much older than

the Sierra Nevada; the latter range is older than the Rocky Mountains; and the Coast Range is still younger.

The normal order of events in the history of a folded range may be interrupted at any stage by renewal or accentuation of uplift, particularly after maturity, causing a revival of stream activity and an increase in ruggedness of relief. Even after a range has been peneplained it may be uplifted and rejuvenated with establishment of a new cycle of erosion. Subsidence would directly cause a lowering of the range and a slowing down of the process of erosion.

Fig. 341. A mountain ridge carved out of strongly tilted strata. Near Banff, Alberta, Canada. (Photo by Miss A. A. Heine.)

Rate and date of folding. It must be clearly understood that the folding and uplift process of a great body of strata into a large mountain range is very slow, generally requiring hundreds of thousands or even some millions of years. As compared to the long eons of known geological time, the active process of folding does, however, take place within a comparatively short time. It is usually much less than the time necessary for the deposition of the strata. Great mountain ranges have been formed by folding of rocks at various times and, in many places, during geological time. Such mountain-making (orogenic) disturbances are commonly called "revolutions." Thus, in North America, since the opening of the Paleozoic era, some of the most important

orogenic disturbances have been as follows: Taconic Revolution, in western New England and southwestern New York, toward the close of the Ordovician period (see table, Chapter I); Appalachian Revolution toward the close of the Paleozoic era; Sierra Nevada Revolution toward the close of the Jurassic period; Rocky Mountain Revolution toward the close of the Cretaceous period; and the Coast Range Revolution at the close of the Tertiary period.

How is the date of a folded range determined? Two principles are involved. First, it is necessary to determine the geological age of the latest (youngest) strata involved in the folding. The folding must

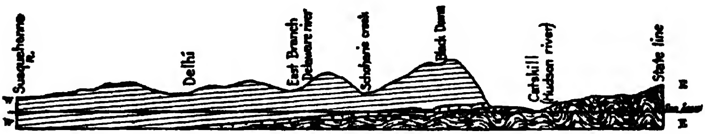

FIG. 342. Structure section illustrating the date of folding of strata. The folded rocks are Ordovician and older; the nonfolded rocks are Silurian and Devonian. Catskill Mountains, New York.

have occurred *after* the deposition of such strata. Second, it is necessary to determine the geological age of the oldest (lowest) nonfolded or less folded strata resting (by unconformity) upon the folded rocks. The folding of the underlying rocks must have taken place *before* the deposition of the overlying strata (Fig. 342). To use a concrete example, we know that the Appalachian Revolution occurred at about the close of the Paleozoic era because the youngest folded strata are of very late Paleozoic age, whereas the oldest nonfolded strata, resting upon the folded rocks, are of early Mesozoic age.

Cause of folding. The great rigidity of the earth and the passage of earthquake waves through it indicate strongly that it is solid, or essentially so, to a depth of about 2000 miles, and that the large central (core) portion is somehow different, being probably liquid under very high temperature and pressure conditions. Mountain-folding, with its accompanying crustal shortening, is quite certainly produced by lateral pressure within the earth's crust. It seems to be a well-established fact that the earth has been a shrinking body for long ages of geological time. It also seems to be clear that orogenic, or mountain-folding, forces are somehow caused by earth contraction with its resultant stresses and strains in the shell (or crust) of the earth.

The ultimate cause of earth contraction is not known. It may be due to loss of heat from the interior, or to the force of gravity whereby the crust is pulled nearer and nearer the center of the earth, or to changes in atomic constitution, or to other causes. Just how earth contraction produces folding of rocks to form mountain ranges is not yet definitely known. Only a few brief suggestions will be given. The whole matter is complicated by the facts that the crustal portion of the earth is a heterogeneous mass, and that folding of strata has been local-

FIG. 343. A mountain ridge carved out of vertical strata. Near Banff, Alberta, Canada.

ized both in place and time, that is, folded mountains have formed in various regions and in various parts of geological time.

"The abundant evidence of compression found in most mountain regions implies a shrinkage of the earth to a smaller volume. Numerous agencies have been recognized as contributory to this result, among which are cooling, crystallization, progressive condensation under gravitative influence, molecular and subatomic changes, redistribution of internal heat, and perhaps other causes. Gravitative readjustments take place periodically, as a result of accumulating stresses, rather than continuously. The origin of igneous magmas is closely interwoven with these readjustments, and the movements or intrusions of these magmas play an important part in the distribution of temperatures, and in the application of localization of stresses within the earth's crust. The net

result of these various agencies, so far as mountain-building is concerned, is to produce the compressive stresses that in turn produce the folding and overthrusting" (G. R. Mansfield).

According to one view, conduction of heat from the interior portion of the earth outward would cause "a severe crowding of the outer (crustal) zone upon itself in shrinking to fit the deep interior as it loses heat and shrinks." Local, relatively weak zones of the crust would yield to such crowding action by crumpling and folding.

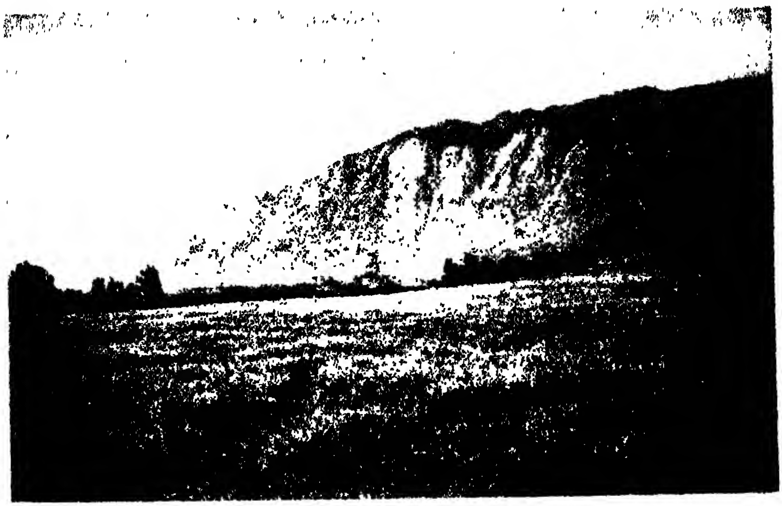

FIG. 344. Part of a great, somewhat eroded fault scarp several thousand feet high. Wasatch Mountains, near Ogden, Utah.

Another view is that the crowding action of the crust upon itself may be largely or wholly due to the force of gravity which acts powerfully to cause all material of the lithosphere to move toward the center of the earth. Weaker portions of the crust would, therefore, be folded.

According to the doctrine of *isostasy*, the lithosphere is in a condition of approximate equilibrium at any given time. The lithosphere consists of heterogeneous material varying considerably in density. Those segments of the lithosphere which consist of heavier material are lower because they are drawn down more powerfully by gravity toward the center of the earth, while the lighter segments stand out in relief. The two grand continental segments and the two grand oceanic segments are thus explained. These grand segments are, in turn, believed

to be subdivided into smaller segments of varying sizes and densities. According to the theory of isostasy, earth segments of comparable surface areas tend to contain the same weight of material irrespective of volume. If, through erosion, much material is transferred from a higher, lighter segment to a lower, heavier segment, rock material at great depth will flow from the heavier segment into the lighter, thus restoring the equality of material in the segments. The heavier segment thus sinks, while the lighter segment rises. Relatively near the surface, material cannot be transferred readily from the heavier to the lighter segment, and so the crowding action of one against the other may cause there a crumpling or folding of the rocks. In harmony with this view is the fact that many great folded ranges have formed along the general borders of the continental and oceanic segments.

According to another explanation, much folding of strata is ascribed to intrusion of great volumes of magma into the earth's crust. Igneous activity is certainly a common accompaniment of mountain folding, and its crowding (or shouldering) action has been deemed by some a sufficient cause of crumpling of adjacent strata.

Fault-block Mountains. Many mountain ranges are caused either partly or wholly by faulting, whereby great earth blocks are made to

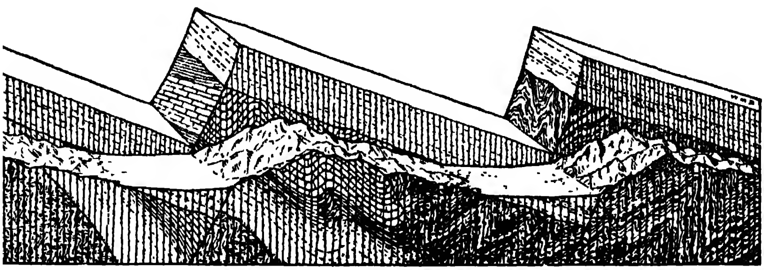

Fig. 345. Diagrams illustrating fault-block mountains. Horizontal strata were laid down upon the peneplained surface of strongly folded strata. Then a series of tilted block mountains developed, and the blocks were much modified by erosion and deposition. Back diagram represents the blocks in potential noneroded form. (After W. M. Davis.)

stand out in relief. Such blocks are often tilted, and they are called *fault-block mountains,* or simply *block mountains.* Tilted block mountains are developed typically in southeastern Oregon (Fig. 356) where a series of them from 10 to 40 miles long and 1000 feet or more high have their fault scarps affected only slightly by erosion. Many of the

north-south ranges of eastern California, Nevada, and Utah are block mountains considerably modified by erosion. The bold western face of the Wasatch Range of Utah is a moderately eroded fault scarp about a mile high and many miles long (Fig. 344). Grandest of all in the

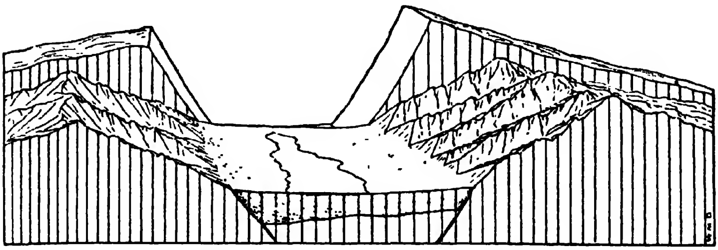

FIG. 346. Diagram showing two tilted block mountains with a fault-block (graben) valley between them. The front diagram shows the blocks modified by erosion and deposition. (After W. M. Davis.)

United States, however, is the Sierra Nevada Range which is a single, great, tilted, fault block over 400 miles long and 60 to 75 miles wide. A somewhat eroded, very steep, fault scarp, ranging from a few thousand feet to two miles high, sharply bounds the range on the east side,

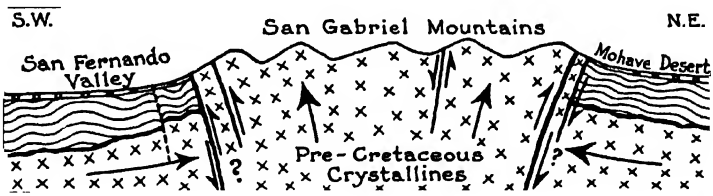

FIG. 347. Diagrammatic structure section to show how the western San Gabriel Mountains of California have been squeezed up as a fault block (horst). Arrows indicate directions of forces involved. The block of crystalline (igneous and metamorphic) rocks is flanked on each side by thick bodies of disturbed strata. Vertical scale is much exaggerated. Length of section about 55 miles.

while a long, relatively gradual slope forms its western side (Fig. 203). A portion of the Rhine Valley of Germany is a sunken fault block lying between two tilted fault blocks—the Vosges and the Black Forest. Fig. 346 illustrates this principle.

Volcanic Mountains. We have already learned that many mountain peaks, often of great height, have been built up by accumulations of igneous materials around the vents of volcanoes. A few among the many well-known examples are: Lassen Peak, California (over 10,000 feet high); Mount Shasta, California (over 14,000 feet); Mount Rainier, Washington (over 14,000 feet); Cotopaxi in Ecuador (nearly 20,000 feet); and Mauna Loa in Hawaii (about 30,000 feet, as measured from the sea bottom on which it stands).

FIG. 348. A great volcanic cone somewhat modified by erosion. Mt. Rainier Washington, as seen from the west. (From a National Park Service Bulletin.)

Not only individual peaks but also whole mountain ranges may be built largely or wholly by volcanic action. Thus the Cascade Range, extending for hundreds of miles from northern California through Oregon and Washington, is to a considerable extent a volcanic range whose once greatest centers of activity are marked by conspicuous cones like Mounts Lassen, Shasta, Pitt, Hood, St. Helens, Rainier, and Baker. The chain of the Aleutian Islands of Alaska, more than a thousand miles long, is an excellent illustration of a mountain range now being built in the sea by active volcanoes. The Hawaiian Islands represent the highest parts of a great, largely submarine, volcanic range hundreds of miles long.

Laccolith Mountains. Closely related to volcanic peaks in origin are so-called *laccolith mountains,* the principle of which is described in Chapter V. In such cases molten material is forced into the crust of the earth, but instead of reaching the surface, it bulges or lifts the upper

portion of the crust into dome-like forms. Laccoliths are very typically illustrated in southeastern Utah and also in parts of Colorado, Wyoming, and South Dakota, in all of which regions practically horizontal strata have been bulged up by magmas. There they show all stages of erosion from those whose covers are practically intact to others whose igneous cores have been more or less laid bare, with the eroded edges of strata lapping up on their flanks (Fig. 114).

Erosion Mountains. All mountains are, of course, subjected to, and modified by, erosion. In not a few cases, however, mountains may be developed by erosion alone in uplifted regions little, if any, affected by

FIG. 349. A mountainous mass, nearly 3000 feet high, carved out of massive sandstone. Zion Canyon, Utah.

folding, faulting, or igneous activity. Such so-called *erosion mountains* are formed by the erosive sculpturing of plateaus and high plains into high ridges, peaks (or buttes), mesas, and deep valleys. The Catskill Mountains of New York, with their numerous narrow ridges and deep valleys, have been carved out of upraised, nearly horizontal strata simply by erosion. The numerous peaks and pinnacles which rise mountain-like within the Grand Canyon of Arizona (Fig. 350) are really erosion remnants or erosion mountains. Another good illustration of erosion mountains is near the mouth of Zion Canyon, Utah, where mountain peaks several thousand feet high have been carved by erosion out of horizontal strata lying from 4000 to 8000 feet above sea level (Fig. 349).

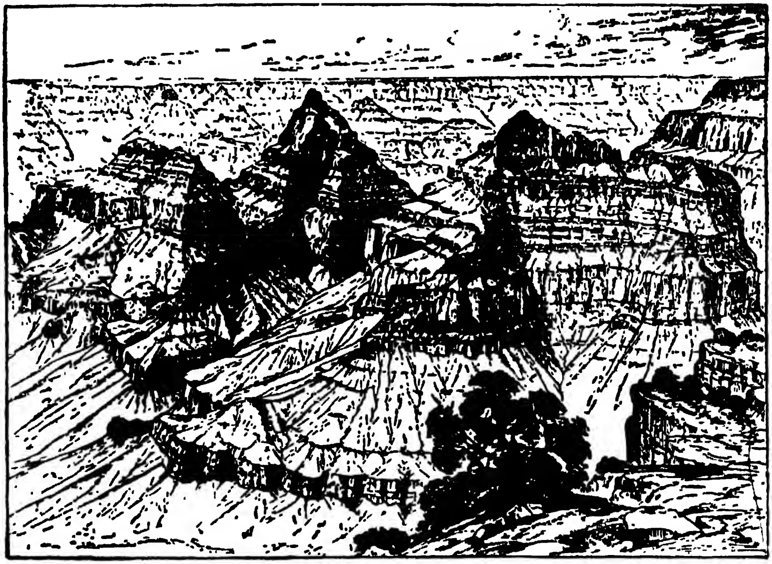

Fig. 350. Mountainous masses several thousand feet high carved out of nearly horizontal strata. Grand Canyon, Arizona. (After Holmes, U. S. Geological Survey.)

Sculpturing and Destruction of Mountains

From the very beginning of its history as a topographic feature, every mountain mass is attacked unceasingly by weathering and erosion which continue to operate during the periods of youth, maturity, and old age. In the course of time every mountain will be leveled by erosion unless it is rejuvenated by some process of igneous activity or diastrophism. Cases of rejuvenation are described under the next heading.

All three of the great erosive agents—water, ice, and wind—are important in the sculpturing of mountains, but water is the most effective. The principles of weathering and of water, ice, and wind erosion, as well as many topographic forms resulting from their action, are described in preceding chapters, and it seems unnecessary to repeat them here in their direct bearing upon mountain sculpturing. Some general effects of the action of running water will, however, be mentioned. Thus a range consisting of well-defined more or less parallel folds will, in maturity or after rejuvenation, be eroded into a system of parallel

ridges and valleys (e.g., the Appalachian Range). A mountain mass consisting of approximately horizontal strata (e.g., the Catskill Mountains) or of igneous or metamorphic rocks with poorly defined, complex structures (e.g., the Great Smoky Mountains) will be carved into a maze of irregular valleys and ridges. Conspicuous valleys will often develop along lines of prominent faults (e.g., the eastern Adirondacks and the Coast Range Mountains). Volcanic and laccolithic peaks will be trenched deeply with valleys radiating from near their summits (e.g.,

Fig. 351. Rugged mountains of irregular shapes carved out of a complicated mass of igneous and metamorphic rocks. They rise 3000 feet above the valley. Front of the San Gabriel Mountains northwest of Pasadena, California. (Photo by J. E. Wolff.)

Mount Shasta). Block mountains will be dissected variously by erosion depending upon the attitude of the blocks and the character and structure of their rocks. Great tilted blocks will, in youth and maturity, have a system of approximately parallel canyons carved out of their long, more gradual slopes, and short steep gorges and canyons in their fault scarps (e.g., the Sierra Nevada Range).

In cold, humid regions, mountains may be sculptured considerably by glaciers which cause development of U-shaped valleys, cirques, and knife-edge ridges (e.g., Glacier National Park, Montana).

The action of wind is an erosive factor usually of moderate importance in mountains in arid regions (e.g., the Great Basin Mountains of Nevada and Utah).

FIG. 352. A three-peak, mountainous mass carved out of granite. The highest summit lies 4000 feet above the valley floor. The influence of large-scale jointing in the production of the three-peak effect is apparent. The Three Brothers, Yosemite Valley, California. (Photo by courtesy of the U. S. Geological Survey.)

REJUVENATION OF MOUNTAINS

The history of many mountain ranges is more or less complex. A range may be born and pass through the mature and old-age stages to extinction (peneplain stage) practically without interruption. It may, or may not, then be rejuvenated by diastrophism or igneous activity. A range may have its normal life history interrupted at some particular

stage, or during more than one stage. The more the great ranges are being studied, the more it is realized that the life histories of many of them are by no means regular and simple. A few examples of interrupted cycles of mountain history will serve to make clear the general principles.

The Rocky Mountains were elevated and more or less folded (Fig. 337) toward the close of the Mesozoic era, at the time of the so-called Rocky Mountain Revolution. This revolution inaugurated a period of

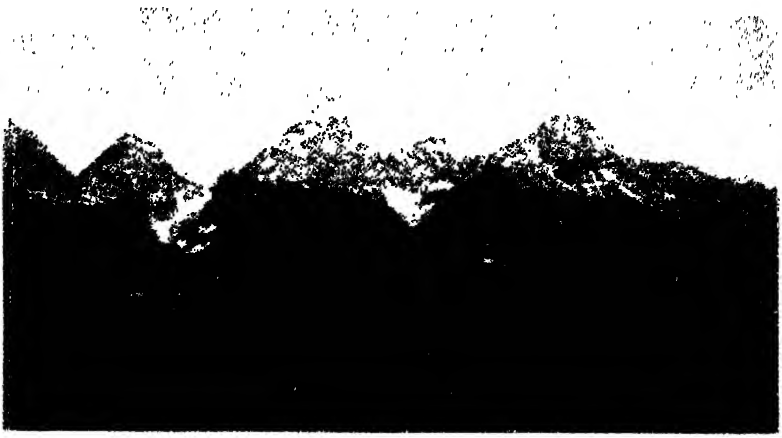

FIG. 353. Rugged mountains carved out of granitic rocks. They rise thousands of feet above the valley in the foreground. Erosive work of both streams and glaciers has been very effective in the development of their ruggedness. Small remnants of the glaciers which formerly flowed through the two U-shape valleys still exist at the canyon heads. Grand Teton National Park, Wyoming.

volcanic activity which affected the mountains considerably during the Tertiary period. In the early Tertiary period there was a notable renewal of folding, especially in Utah, Wyoming, and Montana. Also in early Tertiary time, considerable faulting occurred, particularly the development of the tremendous thrust fault along the eastern side of the Rockies in southern Canada and the northern United States (Fig. 103). In late Tertiary time, and possibly even later, at least one profound movement of uplift accompanied by faulting, but with little or no folding, greatly rejuvenated the Rockies which had been much reduced by erosion. Most of the detailed sculpturing of the mountains has been accomplished since the rejuvenation by rivers and glaciers, during the present (Quaternary) period of geologic time.

The Sierra Nevada Range was severely folded and elevated toward the close of the Jurassic period. Great volumes of granite magma were intruded at the same time. Erosion then held sway until the range was reduced to hills by later Tertiary time. Then a great fault fracture began to develop along the eastern side, and the whole Sierra Nevada fault block has been upraised and tilted into its present position. The many deep canyons, like Yosemite, Kern River, King's River, American River, and Feather River, have been cut into the western slope of the fault block by erosion (Fig. 203).

FIG. 354. A view in the Cascade Mountains. The plateau summit, marked by a nearly even sky-line, represents an old-age surface which has been uplifted thousands of feet and deeply cut into by streams. Gold Creek region of central Washington. (Photo by U. S. Geological Survey.)

The Cascade Range was considerably folded and elevated toward the close of Jurassic time and then eroded. During Tertiary time there were many outpourings of lava in the region. By late Tertiary time the range, particularly in Washington, was eroded to an old age or peneplain condition. In the early Quaternary period there was a great rejuvenation of the range by uplift with warping to form a plateau 4000 to 8000 feet above sea level. That portion of the range which is cut through by the Columbia River was bowed up several thousand feet in the form of a very broad anticline with its axis parallel to that of the range, the river probably maintaining its course during the uplift. The plateau has since been deeply dissected by streams (Fig. 354), and many

large volcanic cones, such as Mt. Shasta, Mt. Rainier, and Mt. Hood, have been built upon the plateau.

The Appalachian (or Eastern Highland) region, extending from New England to Alabama, was more or less severely folded, thrust faulted, and uplifted into great mountains toward the close of the Paleozoic era. This orogeny has been called the Appalachian Revolution. For a long time following the orogeny, the whole region underwent variable changes of level and profound erosion. By Middle Tertiary time the once great mountain region was in a peneplain con-

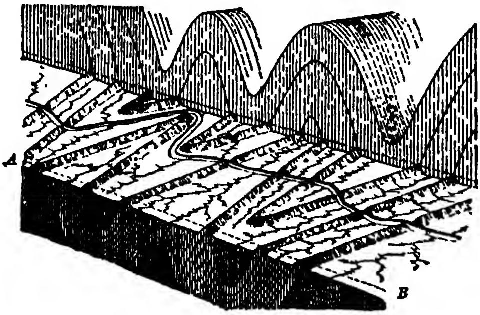

FIG. 355. Block diagram illustrating the history of the middle part of the Appalachian region. (After W. M. Davis.)

dition in its middle portion, and early old age in its northern and southern portions where the rocks were especially resistant to erosion. Since Late Tertiary time the whole worn-down region has been rejuvenated by irregular uplift and warping, but without real folding. The amount of uplift usually varied from 1000 feet to several thousand feet. The existing ruggedness of the range is due to erosion which has operated upon the rejuvenated range (Fig. 355). The system of long, narrow mountain ridges and valleys has been determined by the harder and softer rock formations which follow the strike of the original folds.

Origin and History of Plateaus

Some of the most important modes of origin of great plateaus are the following: (1) By simple uplift with little or no tilting, faulting, or folding, good examples being the plateau of southwestern New York, and the Allegheny Plateau just west of the Appalachian Range; (2) by uplift with tilting and some faulting, but with little or no folding, an

excellent case in point being the great Colorado Plateau; and (3) by the upbuilding of a region by many outpourings of lava, as illustrated by the Columbia Plateau of the northwestern United States. Smaller plateaus (or table lands) may originate by being faulted above the surrounding country, as illustrated by several examples along the southern side of the Adirondack Mountains of New York; or by cutting down the surrounding country by erosion, good examples being the Tug Hill Plateau between Lake Ontario and the Adirondack Mountains, and many mesas of the southwestern states.

Plateaus, as well as mountains, are attacked by weathering and erosion from the very beginning of their history. By maturity, the original, more or less flat surfaces are much trenched and dissected by erosion, giving rise to maximum ruggedness of relief. They are then often called mountains (e.g., Catskill Mountains) instead of plateaus. With continued erosion, plateaus become more subdued in relief, and they are worn down finally to peneplains. As in the case of mountains, so with plateaus, the normal cycle of topographic development may be interrupted by rejuvenation.

Origin and History of Plains

Extensive plains may originate by simple uplift accompanied by little or no tilting, warping, or folding. A good example is the wide Interior Plain of the Upper Mississippi Valley which is nearly everywhere less than 1000 feet above sea level and considerably dissected by erosion.

Great plains may originate by uplift accompanied by notable tilting, as for example the Great Plains area which inclines downward, from an altitude of a mile or more at the base of the Rocky Mountains to a thousand feet or less, within a distance of several hundred miles eastward (Fig. 4). The Great Plains are affected relatively little by erosion.

Plains of great extent may be formed by emergence of a marginal sea bottom, without notable folding, faulting, or warping, either by uplift of the land or by withdrawal of the sea, or by both. Wonderful examples are the Atlantic and Gulf Coastal Plains of the United States (Fig. 4). During the uplift, which was the prime factor in their production, these plains were tilted seaward. They have been dissected considerably by erosion, although many wide, smooth areas remain.

Wide plains which result from long-continued erosion of land are

called peneplains. Many extensive peneplains are known to have developed during geological time, but good examples are rare at present because nearly all those of fairly recent origin have been rejuvenated by uplift. Good examples of such upraised peneplains are those of the Appalachians and southern New England districts. Such uplifted peneplains are best classed among plateaus.

Glacial drift surfaces are often smooth enough for considerable distances to be called plains. A fine illustration is the broad, flat deposit left by the last ice sheet from central to northern Iowa.

Many relatively small plains are floors of extinct lakes which were built up and smoothed off either by deposition of sediment or by mineral matter from solution. An exceptionally fine, large-scale example is the floor of the great glacial Lake Agassiz described in Chapter XIV.

Rivers form flood plains by deposition and lateral erosion, and delta plains by deposition.

Plains may be modified by erosion, diastrophism, vulcanism, or deposition of material. High, or steeply inclined, plains can be affected very profoundly by erosion. High plains, like mountains and plateaus, reach maximum ruggedness of relief during the mature stages of their erosional history. Low plains, by their very position, can be affected but little by erosion. It is an interesting fact that an extensive low-lying plain may last much longer without notable change than a great mountain range.

CHAPTER XIV

LAKES

INTRODUCTIONS

A LAKE is an inland body of standing water. Generally the water is stationary, but it may have a moderate current through it. Lakes always occur where the surface drainage is obstructed. Two necessary conditions are basin-like depressions, and sufficient water at least partly to fill them. Lakes may consist of either fresh or salt water. Fresh-water lakes nearly always have outlets. Some lakes are inappropriately called "seas," as for example the Dead Sea of Palestine.

Lakes vary in size from tiny ponds to bodies of water covering many thousands of square miles, although probably not more than a dozen in the world occupy areas of over 10,000 square miles. Largest of all is the Caspian Sea with an area of 169,000 square miles. The second largest is Lake Superior, covering nearly 31,000 square miles. Next in order of very great size are Lake Victoria-Nyanza in Africa (30,000), Aral Sea in Asia (26,900), Lake Huron (22,322), and Lake Michigan (21,729).

Lakes are known to vary in depth from a few inches to a maximum of 5618 feet in Lake Baikal of Siberia. The Caspian Sea has a depth of at least 3200 feet. Crater Lake, Oregon, with a depth of nearly 2000 feet, is probably the deepest lake in North America. The depth of Lake Superior is 1008 feet. Relatively very few lakes are over 1000 feet deep, and the vast majority of them are less than 100 feet deep.

Most lakes by far lie above sea level at all altitudes up to many thousands of feet. A remarkable case is Lake Titicaca (area, 3200 square miles) in South America at an altitude of 12,875 feet. The highest large lake in the United States is Yellowstone Lake at 7741 feet. Lake Tahoe on the California-Nevada line lies at 6225 feet.

The surfaces of some large lakes lie below sea level, examples being the Dead Sea of Palestine (—1300 feet), the Caspian Sea (—85 feet), and the Salton Sea of California (—249 feet).

The bottoms of a number of large lakes are well below sea level, a few examples being Lake Baikal (—4000 feet or more), Lake Ontario

(—491 feet), Lake Chelan in Washington (—421 feet, and Lake Superior (—402 feet).

Most lakes in humid regions have surface outlets, that is, there is usually sufficient water to cause them to overflow the lowest parts of their basin rims. Such lakes consist of fresh water. In arid regions lakes usually do not have outlets, because of both the scanty volume of water and the high rate of evaporation. Many depressions in arid regions contain no water at all, while others, called *playas,* hold water only temporarily, that is, for greater or less periods after rains. Arid-region lakes with no outlets almost invariably contain salt (or alkaline) water.

As compared to the many millions of years of the known history of the earth, lakes, excepting possibly the largest and deepest ones, are short-lived, most of them exceedingly so. This is because they are merely temporary obstructions to drainage and are soon destroyed by one or more of several processes as explained beyond in this chapter.

Geological Functions of Lakes

Some of the most important geological functions of lakes are the following: (1) Most of the sediments carried into lakes by streams or other agencies settle on their floors. Lakes with no outlets act as perfect settling basins, but even where there are outlets, a great many lakes are effective settling basins. Thus the mighty Niagara River, which drains the four upper Great Lakes, is remarkably clear as it leaves Lake Erie. In other words, practically none of the sediment carried into the four upper lakes leaves them. A remarkable case is Lake Geneva in Switzerland which receives the very muddy Rhone River, and which discharges through a wonderfully clear stream. A very large amount of material accumulates in the lakes of the world each year, but it is much less than the quantity which is deposited on the sea floor. The filtering action of lakes causes the outlet streams to be less effective agents of erosion because, on leaving the lakes, they are lacking in grinding tools (sediments).

(2) Lakes act as storage reservoirs for surface drainage, and so they tend to regulate the volumes of streams which flow out of them, furnishing a check upon destructive floods. The rate of erosion is thus affected because, as we have already learned, streams are usually most effective in their power to erode in times of flood. With slow-moving graded streams, there would be less tendency to develop wide flood

plains, and, as a result of the settling-basin effects of lakes, such streams would deposit less material upon their flood plains.

(3) Larger lakes or numerous smaller ones tend to increase the rainfall and also to keep the temperature of a region more uniform, making summers cooler and winters milder. This is well illustrated in the regions bordering Lakes Erie and Ontario on the south and southeast. Such climatic influences of lakes are of real importance because various geological processes, like weathering and erosion, are thereby affected.

(4) Plants or shell-bearing animals, or both, are abundant in many lakes. Their shells, skeletons, or partly decayed remnants accumulate in the lake basins, thus helping to fill them. Such materials, found in fossil form in old lake deposits, often yield valuable evidence as to the nature of certain phases of the life of the geological times, as far back as millions of years ago, when the particular lakes existed.

(5) Considerable work of erosion is accomplished by waves cutting into parts of the shores of many lakes, the materials eroded being spread over the floors of the basins.

(6) When lakes with no outlets evaporate in arid regions, the various salts held in solution are deposited in layers or beds over the sites of the lakes. There are many examples of such deposits in the arid region of the western United States.

Origin of Lake Basins

Lake basins originate in many ways. Our present purpose is to explain most of the more common, important, and interesting modes of origin of lake basins and to describe briefly some illustrative examples.

Basins Formed by Diastrophism. *By faulting.* When a block of the earth's crust sinks or rises relative to an adjacent block through the

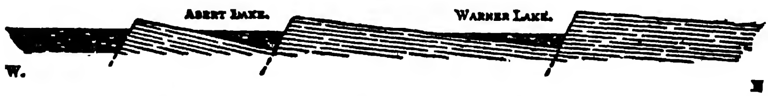

Fig. 356. Structure section illustrating the origin of lake basins by faulting. Abert and Warner Lakes, Oregon. (After Russell, U. S. Geological Survey.)

process of faulting, a troughlike basin often results. There are many examples in the Great Basin region of the western United States. A small basin, now partly filled with water, formed in 1872 as a result of

a sudden renewed movement of 20 to 30 feet along a fault near the base of the Sierra Nevada Range in the Owens Valley of southeastern California. Abert and Warner Lakes of southern Oregon are very typical cases of lakes in basins between tilted fault blocks (Fig. 356).

Great Salt Lake (Fig. 357) occupies the lowest portion of the surface of a vast block of earth which has been depressed thousands of feet by faulting relative to the adjacent Wasatch Range of Utah. This remarkable lake is described beyond under the heading "Salt Lakes."

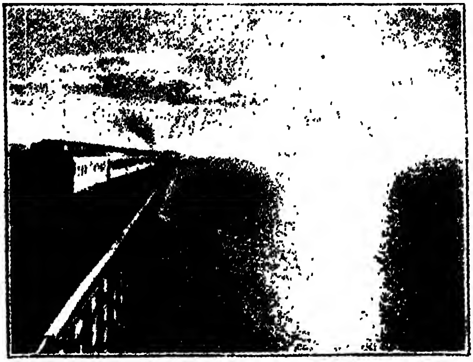

FIG. 357. A view across Great Salt Lake, Utah. (Photo by courtesy of the Southern Pacific Lines.)

Lake Tahoe, on the California-Nevada line (Fig. 358), is a fine example of a lake occupying a basin produced largely by relative down-faulting of an earth block several thousand feet. Lava flows across the outlet of the basin accentuated its depth somewhat. It was once considerably deeper, but cutting down of its outlet by erosion has lowered its level. The lake is 22 miles long and 12 miles wide. Its surface lies 6225 feet above sea level. It has a depth of at least 1645 feet, making it one of the deepest lakes in North America. Its water is remarkably clear and fresh, with an outlet through Truckee River.

A still more remarkable example of a trough-fault basin containing a large lake is the Jordan Valley of Palestine with the Dead Sea in its lowest portion. The Jordan Valley was formed by the settling of a very long, narrow block of earth thousands of feet between two normal faults. Much of it, including the lake, lies well below sea level. The Dead Sea is described beyond under "Salt Lakes."

Lake basins sometimes come into existence by shifting or settling of earth blocks during earthquake disturbances. Mention has already been made of such a lake in the Owens Valley of California. During the violent earthquake of 1811-1812 in the New Madrid, Missouri region, a number of lake basins were formed, the largest being occupied by Reelfoot Lake, Tennessee.

By warping. Warping of the earth's crust through differential movement also has caused the development of lake basins. Part of a river valley may be sufficiently upwarped to act as a dam, causing ponding of the water. Among examples ascribed to such a cause are the basin of Lake Geneva in Switzerland and of Lake Temiskaming in Ontario, Canada. An exceptionally fine, large-scale example caused by warping of a wide area is the great Caspian Sea described beyond under "Salt Lakes." Among other large lake basins, which at least in part owe their existence to warping, are those of the Great Lakes described beyond.

Fig. 358. A view across Lake Tahoe from the Nevada side to California. (Photo by courtesy of Tavern Studio, Tahoe, California.)

By simple uplift. When a portion of the sea bottom is raised into land, without faulting or notable warping, there often are shallow, irregular, basin-like depressions filled with water. The water is at first salty, but, in humid climates, it soon gives way to fresh water. A number of the lakes of the southern half of Florida and of the plains of Siberia are believed to be of this origin. Such lakes are very short-lived because of their shallowness.

Basins Formed by Vulcanism. *Crater lakes.* Numerous lake basins are direct results of volcanic activity. Many of them are simply craters

FIG. 359. A view across the western part of Crater Lake, Oregon, showing Wizard Island—a recent cinder cone. (Photo by courtesy of the Southern Pacific Lines.)

of inactive volcanoes more or less filled with water. Sometimes there are groups of such crater lakes, as in the Auvergne district of France, the Eifel region of Germany, and the vicinity of Rome, Italy. These are all small, but beautiful, lakes.

Many crater lakes also occur in large and small craters of volcanoes in the western United States. Most remarkable of all is Crater Lake in the Cascade Mountains of southern Oregon (Fig. 359). It partly fills a vast hole (caldera), six miles in diameter, which resulted from explosive activity followed by collapse of the upper portion of a lofty volcanic cone. The lake is nearly 2000 feet deep, being probably the deepest in North America. Its surface lies nearly 6200 feet above the sea. Great precipitous walls of rock completely encircle the lake. It has no surface outlet, and yet its water is fresh, probably in part because some of its water may leave by underground passages, and in part because no stream flows into it. The water supply is maintained by rainfall and snowfall.

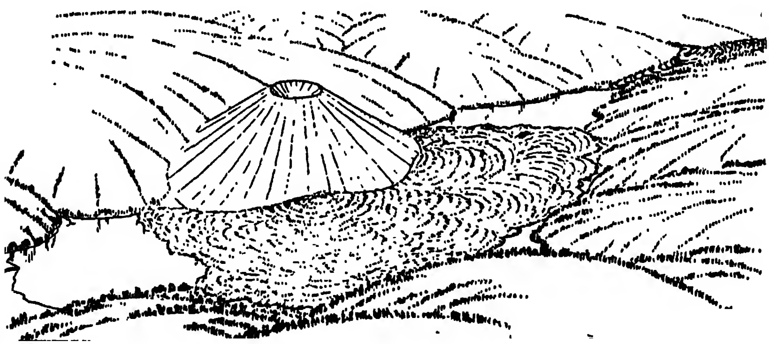

FIG. 360. A sketch showing how a lava flow, at the base of a cinder cone, poured across a lake-filled valley, leaving a part of the lake at each end of the flow. This kind of thing happened in Lassen Volcanic National Park. (After W. M. Davis.)

The vast crater pit formed by the terrific explosions of Katmai Volcano, Alaska, in June, 1912, contains a lake of warm water about a mile in diameter.

Lava-dam lakes. Streams of molten lava may flow across valleys and there cool to form natural dams, causing the valley waters to be ponded. Thus a great flow of lava from Skaptar Jökull, Iceland, in 1783, blocked a large river and a number of its tributaries, with resultant development of lakes.

The famous Sea of Galilee in Palestine was formed by a stream of lava which, in very recent geological time, flowed into and across the Jordan Valley, causing the River Jordan to be ponded nearly 700 feet below sea level. The water is fresh because the river flows through the lake.

A number of lava-dam lakes occur in the Sierra Nevada and Cascade Mountains of the western United States. An interesting example is Snag Lake in Lassen Volcanic Park, California, whose water level is held up by the very recent flow of lava already described as partly filling a valley (Fig. 360).

Basins Formed by Glaciation. *By glacial-drift dams.* Lake basins formed by various processes of glaciation are more abundant than those formed in any other way. Of these, the most numerous by far have resulted from the deposition of glacial débris (moraines) in such manner as to obstruct the drainage of valleys. Many of the 8000 or more lakes in Minnesota, of the thousands in Wisconsin, and of the thousands in New York and New England belong in this category.

Fig. 361. A lake formed by blockading a valley with a morainic dam. Blue Mountain Lake, New York.

The most common types of moraine-dam lakes are finely illustrated in the Adirondack Mountains of northern New York (Fig. 361). Thus the well-known Lake Placid, Saranac Lakes, and Long Lake have their waters ponded by single dams of morainic materials across valleys. They all lie between 1500 and 1900 feet above sea level. The Fulton

Chain of eight lakes in the Adirondacks illustrates a series of lakes which occupy basins formed by a succession of morainic dams across a valley. Lake George in the southeastern Adirondacks is a fine example of a large, long, narrow body of water maintained by two morainic dams across a valley, one at each end of the lake. It is 30 miles long, from one to two and one-half miles wide, and lies in a deep, narrow mountain valley.

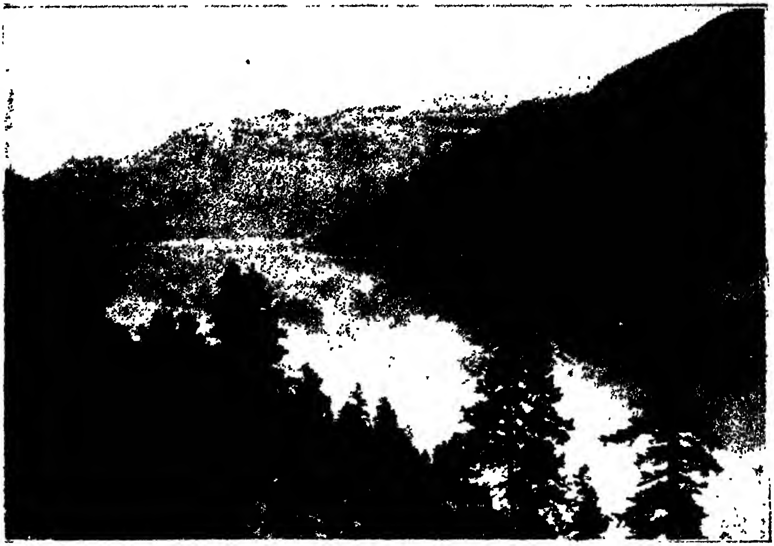

FIG. 362. A view of part of Lake Chelan, Washington. Its basin was produced partly by glacial erosion, and partly by a moraine dam. (Photo by courtesy of the U. S. Reclamation Service.)

Lake Chelan in the Cascade Mountains of Washington is, in regard to length, depth, narrowness, and scenic setting, one of the most remarkable mountain lakes of the United States (Fig. 362). It is 60 miles long, less than two miles wide, about 1500 feet deep, and set in a winding mountain canyon several thousand feet deep. First a river carved out most of the canyon. Then a great valley glacier plowed through the basin, deepening it many hundreds of feet, and leaving a heavy morainic accumulation across its lower end, thus still further increasing the depth of the lake basin.

The three famous lakes of northern Italy—Como, Garda, and Maggiore—occupy deep, narrow, steep-sided mountain valleys on the southern slope of the Alps. A big glacier flowed down each of these valleys during the Ice Age and spread out part way upon the Italian plain. Great terminal moraines were formed around the ends of the glacial lobes, and, since the melting of the ice, the moraines have acted as dams across the ends of the mountain valleys, ponding the waters far back in them. Each of these lakes is over 1000 feet deep. Similar to these Italian lakes in general character, although not so large and deep, but like them in origin, are a number of long, narrow lakes of Glacier Park, Montana.

By ice dams. Glaciers may blockade valleys and thus cause ponding of waters. Lakes of this kind occur in Greenland, Alaska, and the Alps. An example is a small lake formed where the Great Aletsch Glacier in the Alps slowly flows past the mouth of a tributary valley, causing a ponding of water in the latter. A great glacier flows into the Copper River of Alaska, ponding the water there.

Existing ice-dam lakes are not common, and few, if any of them, are large. During the Ice Age, however, thousands of them formed and lasted only as long as the ice dams existed. Some of them were of vast extent, vaster in fact than any existing lakes, with the possible exception of the Caspian Sea. A few large-scale examples will be briefly described.

When the great ice sheet was retreating from northern New York, waters hundreds of feet deep and many miles long were ponded in the Mohawk Valley between the great walls of ice—one in the eastern and the other in the western part of the valley. The lake waters lowered as the ice retreated, and they finally drained away completely.

One of the largest of all known ice-dam lakes has been named Lake Agassiz. It occupied the Red River Valley region of Manitoba (including Lake Winnipeg), North Dakota, and Minnesota. It attained a maximum length of about 700 miles, and a width of over 200 miles, when it covered 110,000 square miles, or considerably more territory than all of the Great Lakes. The lake formed because the northward drainage into Hudson Bay was blocked by the front of the retreating ice sheet during a late stage of the Ice Age. The outlet of this vast lake was southward into the Mississippi River.

The Great Lakes constitute the most remarkable chain of big lakes in the world. They cover about 95,000 square miles. Their history is too complicated and eventful to be more than suggested in a very brief

account. They did not exist before the Ice Age. Their basins are the results mainly of extensive deposition of morainic materials on their southern sides and notable down-tilting of the land toward the north, during the Ice Age. The broad valleys which formerly occupied the sites of the various basins were also more or less deepened by glacial erosion. Our present interest is, however, a very brief consideration of the wonderful systems of ice-dam lakes which developed during the retreat of the great glacier from the Great Lakes basins. When the front of the ice sheet withdrew far enough northward to free the

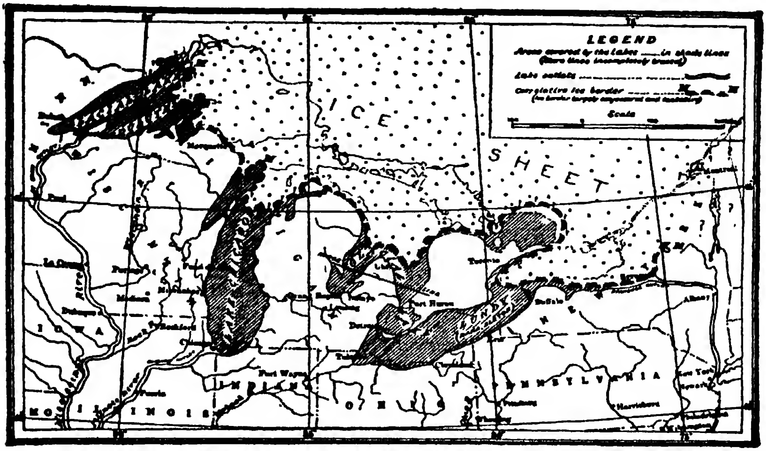

FIG. 363. A three-lakes stage of the history of the Great Lakes during the retreat of the great ice sheet. (After Taylor and Leverett.)

southern end of the Michigan basin and the western end of the Erie basin, small glacial lakes developed against the ice walls in those basins. The first-named drained southwestward through the Illinois River, and the other southwestward through the Wabash River—both into the Mississippi. These lakes enlarged as the ice retreated somewhat farther, and the eastern one drained across Michigan into the western one which, in turn, drained through the Illinois-Mississippi Rivers. At a later stage, when fully half of the Great Lakes basins was freed from the glacier, three large, independent lakes lay against the ice wall (Fig. 363). One of these (Lake Duluth) filled the western half of the Superior basin, with outlet through the St. Croix-Mississippi Rivers. Another filled most of the Michigan basin, with outlet through the

Illinois-Mississippi Rivers. The third lake occupied the Erie Basin and the southern end of the Huron basin, with outlet eastward through the Mohawk-Hudson Valleys of New York. During a still later (Algonquin-Iroquois) stage, the Great Lakes assumed nearly their present condition, though they were somewhat larger, and they all drained through the Hudson-Mohawk Valleys of New York because the St. Lawrence Valley was still filled with the glacier (Fig. 364). Finally

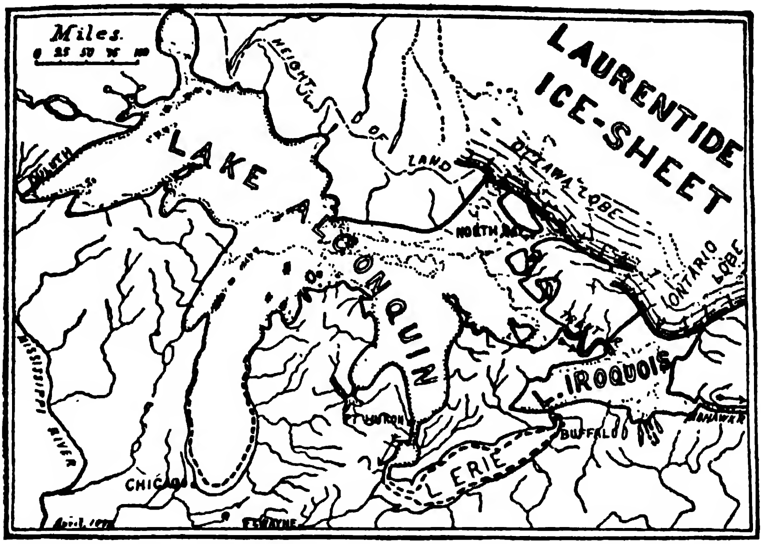

FIG. 364. The Algonquin-Iroquois stage of the Great Lakes history. (After Taylor.)

the ice disappeared from the St. Lawrence Valley, and soon the present-day conditions obtained. Certain earth-crust movements accompanied the various changes mentioned, so that the Great Lakes basins are also in part of diastrophic origin.

By glacial erosion. A considerable number of glacial lake basins have been eroded or excavated by the direct action of flowing ice. Small *rock-basin lakes* of this kind, usually not more than ponds, often occur in the bottoms of the cirque basins at the heads of valleys formerly occupied by valley glaciers, because the excavating power of such glaciers was there especially effective (Fig. 365). Less often, rock basins have been excavated by glaciers farther down their valleys. Valley-glacier, rock-basin lakes are numerous in parts of the Sierra Nevada, Cascade, and Rocky Mountains (Fig. 366), and also in the

high mountains of Europe. Much of the depth of Lake Chelan, Washington, is a result of glacial erosion (Fig. 362).

Other rock basins, including some large ones, were produced by the erosive action of the great ice sheets during the Ice Age. Some of the numerous lake basins of Ontario, Canada, quite certainly belong in this category. Ice erosion was especially effective in that region, while glacial deposition predominated from the Great Lakes southward. A large lake recently assigned to this class of rock-basin lakes is Lake Athabaska (area, 2800 square miles) in central-western Canada.

FIG. 365. A small rock-basin lake in the bottom of the cirque on the eastern side of Longs Peak, Colorado. (Photo by W. T. Lee, U. S. Geological Survey.)

Many glacial lake basins owe their existence to a combination of erosion and deposition. The Great Lakes basins have already been mentioned as belonging in this category. Among many others are the Finger Lakes which form a remarkable group in central-western New York.

By irregular deposition of glacial drift. Many ponds and small lakes occupy depressions which have resulted from irregular deposition of glacial débris (drift). Such basins are merely depressions in the surface of the drift. They are common in the upper Mississippi Valley, New York, and New England, especially in association with the many recessional moraines. They differ from typical morainic dam basins not only in that they are completely surrounded by drift, but also in that they very commonly developed on flat, or only slightly hilly, land.

Ponds and small lakes may occupy depressions formed by the melting of large, isolated blocks of ice which have become buried under sediments. Masses of ice detached from a glacier may have been covered by morainic

material left by the ice; or such masses may have been buried under material washed from the glacier (as in valley trains and outwash plains); or icebergs stranded in glacial lakes may have been buried under sediments carried into the lakes by streams. Ponds and lakes in such depressions are called *pit* or *kettle lakes.* They are most strikingly shown on otherwise nearly level, loose, extensive deposits which mark the sites of former glacial lakes. When such a surface is characterized by many kettle holes, some with and some without water, it is called a *pitted plain.*

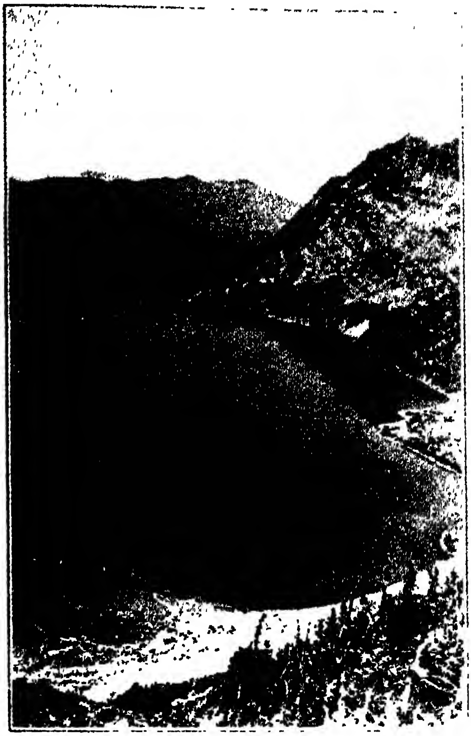

FIG. 366. A rock-basin glacial lake. Lake Ellen Wilson, Glacier National Park, Montana. (Copyright photo by R. E. Marble.)

Basins Formed by Stream Action. *By flood plain development.* We have already learned that graded and nearly graded rivers tend to wander in meandering loops over their flood plains, and that the necks of such loops are often cut across, leaving *oxbow lakes* like those so wonderfully exhibited on the flood plain of the lower Mississippi River.

Shallow basins often result from uneven deposition of the flood-plain sediments, especially in the spaces between the natural levees of the main streams and their tributaries.

By delta growth. As a result of uneven deposition of sediment by the network of distributaries on a delta, certain shallow basins are completely surrounded by the deposits, and thus converted into so-called *delta lakes.* A fine large-scale example is Lake Pontchartrain in Louisiana.

By alluvial cones. An alluvial cone or fan formed by a tributary stream may be built far enough out into its main stream (or valley) to obstruct the drainage of the latter, causing a ponding of the water. A good case in point is Lake Pepin which lies between Minnesota and Wisconsin. Much sediment carried by the Chippewa River into the Mississippi has there caused a ponding of the latter.

By raft blockades. In Chapter VIII mention is made of stream obstruction and deflection caused by so-called *rafts* or *jams* of trees and logs formed in rivers. The growth of such a raft upstream for many miles in the Red River of Louisiana so obstructed its tributaries as to develop a remarkable series of small and large lakes along them.

By waterfall erosion. Small lakes are sometimes found in abandoned stream courses, particularly where waterfalls have excavated so-called "plunge basins" at their bases. Fine examples of *plunge-basin lakes* are Jamesville Lake near Syracuse, New York, and near Coulee City, Washington, where large rivers once flowed.

Basins Formed in Other Ways. Brief mention will be made of some of the other modes of origin of lake basins, with examples.

By waves and shore currents. When the mouths of embayments of either sea or lakes are closed by the growth of bars or barriers through the action of either shore currents or waves, or both together, lagoonal lakes result. Many examples occur along the Atlantic Coast from Long Island southward, around the borders of the Great Lakes, and along the Pacific Coast of the United States (Fig. 308).

By wind. Wind action often piles the materials of bars and barriers higher, thus causing them to be more effective dams where they close embayments of sea or lakes. Wind-blown sand may block streams locally, causing ponding of their waters. This has often happened along the southwestern coast of France. Depressions in sand dune areas sometimes contain water. Wind erosion may, under exceptional conditions, excavate basins in soft rock materials, as in parts of Argentina.

By solution. When sink holes are sufficiently obstructed by rock débris at their bottoms they may contain ponds or small lakes. Good examples occur in the northern half of Florida and in Kentucky.

By landslides. Lakes are sometimes formed where landslides obstruct the drainage in valleys and canyons, particularly in regions of high relief. A good example is in the Kern River Canyon of the southern Sierra Nevada Mountains. In 1892 a great landslide blockaded the upper Ganges River in India, causing a lake five miles long and

hundreds of feet deep. The lake disappeared in about two years by a giving way of the dam.

Salt Lakes

Origin. Salt lakes are far less common than fresh lakes. They never have outlets. They almost invariably exist in arid regions, particularly in *interior drainage* regions, like the Great Basin area of the western United States, from which no streams flow into the sea. In such regions the intake (precipitation and inflow) is often not sufficient to cause the lakes to overflow the lowest points of their basins to form outlets. With increase in dryness of climate, a fresh lake may, therefore, become a salt lake because the outlet is sooner or later abandoned, and mineral matter, carried in by streams, steadily accumulates in the water. Great Salt Lake, Utah, is one of the best-known examples belonging in this category. A salt lake may, under certain conditions, become a fresh lake. Thus Lake Champlain, which became detached from the Gulf of St. Lawrence by uplift of the land, was a salt lake at first, but the salt has since been rinsed out through the outlet stream. Such a body of water, which was once connected with the sea, is called a *relic lake.*

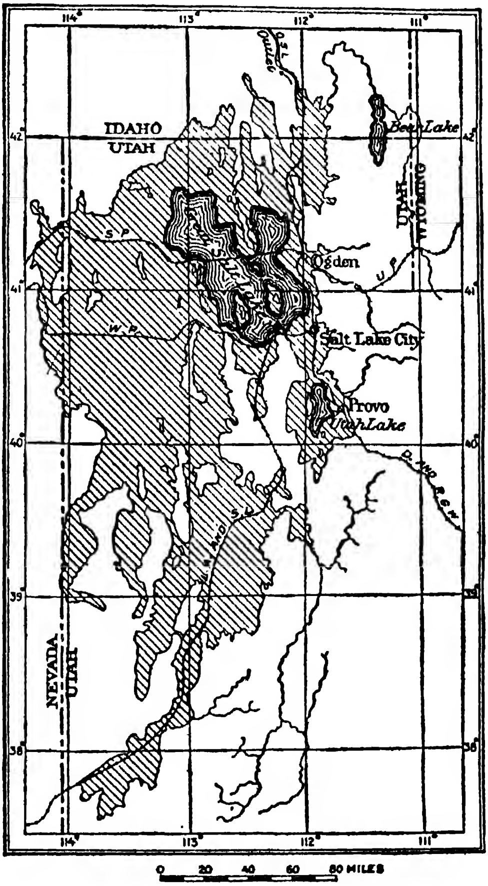

Fig. 367. Map of Great Salt Lake and its vast fresh-water ancestor, called Lake Bonneville (shaded). (After U. S. Geological Survey.)

Salt lakes may, in short, be formed in two ways, namely, (1) by accumulation of saline matter in lakes with no outlets, and (2) by cutting off arms of the sea either by diastrophism or by deposition of sediment, particularly in the form of a delta. Examples illustrating these principles will now be briefly described.

Examples. Great Salt Lake, Utah, is a fine example of a salt lake not only whose saline matter has accumulated by concentration through excessive evaporation, but also whose ancestor was a fresh lake. As already mentioned, it occupies the lowest portion of the surface of a vast down-sunken fault-block in the Great Basin region (Fig. 357). It covers nearly 2000 square miles, and its surface lies 4200 feet above sea level. It is remarkably shallow, the greatest depth being only about 40 feet. It is nearly five times as salty as the ocean, that is, it carries about 18 per cent of saline matter in solution. It contains several billions of tons of common table salt, and hundreds of millions of tons of salts of soda, magnesia, potash, lime, etc. Very briefly stated, the history of the lake is as follows: when the climate was moister, the vast basin, now only partly occupied by the lake, was filled to overflowing with fresh water (Fig. 367). This great lake, called Lake Bonneville, was about two-thirds the size of Lake Superior, and its outlet was northward into Snake-Columbia Rivers. Lake Bonneville had a maximum depth of over 1000 feet. As the climate became drier, evaporation exceeded intake, and the outlet was abandoned. The water level fell, though not uniformly, and the lake became more and more salty by concentration of saline matter carried in by streams. Great Salt Lake is but a shrunken remnant of its vast ancestor. Many shore-line features, such as bars, beaches, deltas, and wave-cut cliffs, marking various levels of the lowering waters of Lake Bonneville, are wonderfully preserved around the sides of the basin (Fig. 368).

FIG. 368. High-level shorelines of Lake Bonneville, Utah. (After Gilbert, U. S. Geological Survey.)

The great Caspian Sea is a fine example of a large arm of the sea

cut off by uplift of land. It is hundreds of miles long, and it covers an area larger than the state of California. It lies 85 feet below sea level, and its greatest depth is over 3000 feet. Both the composition of the salts in solution and the nature of the animal life in its water point to its former connection with the ocean. The Caspian Sea formerly connected with the Black and Mediterranean Seas across southern Russia, and it was cut off by a broad, gentle upwarp of the land there. A remarkable fact is that this vast body of water with no outlet now

FIG. 369. A briny lake rich in sodium salts in a desert basin. The white areas are covered with salts left by the retreating brine. A large fresh-water lake once occupied this basin. Deep Spring Valley, California.

contains less salt (less than one per cent) than when it was connected with the ocean. This is owing to the fact that water from the Caspian Sea steadily flows through narrow openings into embayments along its sides, particularly the Gulf of Kara, situated where the climate is arid. Great evaporation of the water in the embayments causes the influx of water from the great lake into them. The water of the embayments evaporates, but the salt remains, and even accumulates in beds. Thus the original salt of the Caspian Sea is slowly being removed, and the large quantities of mineral matter brought in yearly by the rivers on the north are not enough to increase the salinity of the lake.

The Salton Sea occupies the lowest part (Salton Basin) of a great

desert depression extending through the Coachella and Imperial Valleys of southeastern California to the head of the Gulf of California in Mexico. The Salton Basin portion of the depression has subsided to

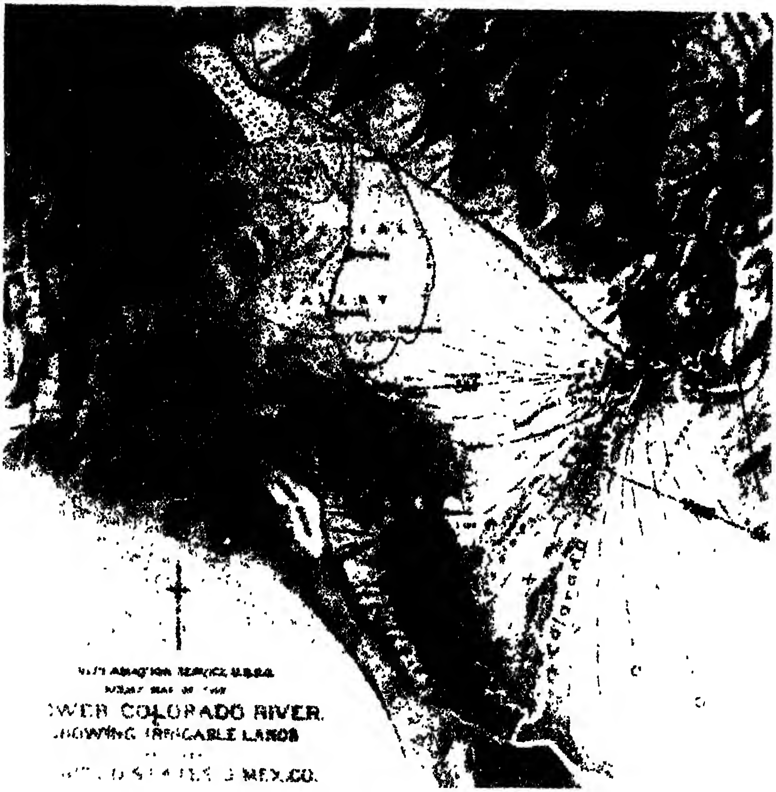

FIG. 370. Map of parts of southeastern California and northwestern Mexico. The radiating broken lines show the great fan-shaped alluvial deposit built by the Colorado River across the desert depression which extends many miles northwestward from the Gulf of California. The dotted area shows the maximum (1907) stage of the Salton Sea when its surface was 198 feet below sea level. The whiter portion, marked "Salton Sink," is the present approximate size of the Salton Sea. (After Reclamation Service and U. S. Geological Survey, with additions by the author.)

below sea level and it would now be occupied by Gulf of California water but for the fact that the Colorado River has built an extensive alluvial fan deposit across the depression, thus separating the Salton Basin from the Gulf of California (Fig. 370). The Colorado River

because of its shifting course on the fan, has, from time to time, poured water into the Salton Basin. The last time this happened the record of shore lines and shells shows that a fresh-water lake, covering 2000 square miles, filled the basin to overflowing. This water evaporated leaving only a desert basin mostly below sea level with salt beds in its lowest portion. Between 1904 and 1907, as a result of an accident to the headgate of a great irrigation canal, much of the Colorado River flowed into the basin, and formed the Salton Sea, covering 450 square miles, in its lowest portion. Since 1907 the body of water has diminished in size considerably by evaporation. Its surface is now nearly 250 feet below sea level, and it contains approximately one per cent of salts in solution.

The Dead Sea of Palestine lies in the lowest portion of the Jordan Valley which was formed by the sinking of a long, narrow block of the earth's crust between two nearly vertical, parallel faults. It covers an area of about 500 square miles; its greatest depth is about 1300 feet; and its surface lies about 1300 feet below sea level, making it the lowest lake in the world. Approximately 24 per cent of salts, chiefly chloride of magnesia and common salt, are in solution in its water. The Dead Sea is but a remnant of a once much larger (fresh-water) lake which had an outlet to the south. As the climate became drier, excessive evaporation caused the water level to lower more than a thousand feet, that is, to the present level of the Dead Sea. The salts in solution have been concentrated from the fresh water brought in by the streams, especially by the Jordan River.

Among other interesting salt lakes, mention may be made of Mono Lake, California, over 6000 feet above the sea, and rich in soda and salt; Owen's Lake, California, nearly 4000 feet above the sea, and very rich in soda; Aral Sea, Siberia, a very large lake (area, nearly 27,000 square miles) only a little above sea level, and only slightly salty; Lake Van, Armenia, which contains 33 per cent of salts in solution; and Assal Lake in eastern Africa, the surface of whose salt water lies 600 feet below sea level.

Lake Erosion and Deposition

Sea and Lake Erosion and Deposition Compared. Lake erosion is, in almost every way, like sea erosion, except that the effects produced in lakes are usually less conspicuous because they are smaller, and the waves and undertow, which do nearly all the erosive work, are less

powerful. In large lakes, however, the resulting features of lake erosion, such as wave-cut terraces, cliffs, caves, coves, stacks, and arches, are practically identical with similar features produced by sea erosion.

Deposition in lakes is, in many respects, also like deposition in the sea. Thus beaches, barriers, bars, deltas, and wave-built terraces are essentially the same whether formed in sea or lakes. These have all been considered under "Marine Deposits" in Chapter XI. There are, however, no deposits in lakes really comparable to deep-sea deposits, with the possible exception of some accumulations of shells of certain tiny plants (diatoms). More or less land-derived sediments carried in by streams accumulate over the entire floors of most lakes. Delta-growth in lakes is particularly strong because of both absence of very appreciable tides and the usual lack (except in some very large lakes) of very powerful wave and undertow action.

Cycles of shoreline development are also essentially the same for lakes as for the sea.

A deposit rather characteristic of some fresh lakes is *marl,* which is a light gray mixture of carbonate of lime (often in the form of shells) and clay. It is often in beds 5 to 20 feet thick.

More or less decaying vegetable matter accumulates in many lakes and swamps in humid-climate regions. Such material often forms rather extensive beds of so-called *peat.*

Chemical Deposits in Lakes. Chemical deposits (various salts) are seldom of importance in fresh lakes. In certain fresh lakes fed by streams rich in dissolved carbonate of iron, the soluble material may, on entering the lakes, become oxidized to the insoluble *limonite* and be deposited on the lake floor. In some lakes of Sweden there is enough limonite of such origin to be of commercial value as an ore of iron. Carbonate of iron may, under exceptional conditions, precipitate in fresh lakes.

The chemical precipitates of lakes are, of all lake deposits, probably the most interesting, characteristic, and important. When, through evaporation, a salt lake shrinks, the minerals in solution become more and more concentrated until deposition of certain of them begins. With continued evaporation to dryness, all minerals in solution are deposited (Fig. 374). Some of the most abundant of many minerals contained in the waters of salt lakes are common salt (halite), sulphate of lime (gypsum), sulphate of soda (Glauber Salt), sulphate of magnesia (Ep-

som Salt), chloride of magnesia, and carbonate of lime (calcite) (Fig. 369).

The kinds and relative amounts of substances in solution depend largely upon the nature of the rocks surrounding the lake basins because (except in certain relic lakes) they are washed out of those rocks and carried into the lakes by streams. Thus Great Salt Lake is very rich in common salt because the surrounding rocks are mainly strata of marine origin containing original sea salt. Mono Lake, California, is rich in carbonate of lime and soda because the surrounding rocks are mainly igneous which, on weathering, yield carbonates.

During the desiccation of a salt lake, dissolved substances are deposited in the order of their insolubility. Thus if four substances, carbonate of lime, sulphate of lime, common salt, and chloride of magnesia, are in solution, they will precipitate in the order given, the last-named remaining in solution the longest because it is very highly soluble. The entry of flood waters during a rainy season or a series of rainy seasons may dilute the water enough to check precipitation of mineral matter from solution, and land-derived sediment will accumulate on the lake floor instead. It is, therefore, readily seen why alternating layers of clay or sand and one or more salts are often found in old lake deposits. Extensive deposits of soda and borax mark the sites of some former lakes, as in parts of Death Valley, and in other basins of southeastern California. Carbonate of lime is now accumulating in Great Salt Lake and lime deposits of curious shapes and large volume occur in Pyramid Lake, Nevada, and in Mono Lake, California. The extensive salt fields just west of Great Salt Lake were left by the retreating waters of the lake.

Destruction of Lakes

By Filling with Sediments. This is one of the most important methods of lake destruction. Some one has said that "rivers are the mortal enemies of lakes." All surface waters, especially streams, flowing into lakes carry more or less sediment with them. Most of the sediment accumulates on the floors of the lakes because the latter are such excellent settling basins as already explained (Figs. 371 and 372). Lake basins may, by this process alone, be completely filled, and the lakes destroyed.

One of the most striking features of the sediment-filling process is delta-growth of the coarser material at the mouths of the tributary streams. As the deltas build out, the lake waters are, of course, dis-

placed. A few examples may be mentioned. Streams entering the heads of both Seneca and Cayuga Lakes in central-western New York have

FIG. 371. A filled lake basin at an altitude of 9000 feet near Pando, Colorado.

FIG. 372. Mountains almost buried under the sediments of former Lake Bonneville, Utah. (After Gilbert, U. S. Geological Survey.)

built delta plains into each about three miles long, and one mile wide. The Rhone River has built a delta 20 miles long into Lake Geneva, one mile of it during the last 1900 years. Finer sediments are, of course, more widely distributed over floors of lakes.

The materials eroded by waves along lake shores are mostly deposited in the lakes. Although such erosion enlarges the areas of lake basins, nevertheless their waters are, on the average, steadily made shallower because most of the material eroded and deposited comes from well above the lake levels.

By Filling with Organic Remains. In humid, temperate-climate regions, many small lakes have been, and are being, destroyed by accumulations of vegetable matter, shells, etc. Plants usually grow in great profusion in the shallow, border portions of lakes. As the plants die their remains accumulate to form bogs which, in many cases, have

FIG. 373. A small lake partly filled with vegetation. Tall grasses encroaching in the water, and brush and trees growing on a bog-filled portion. Horseshoe Park, in Rocky Mountain National Park.

encroached from all sides until lakes have been completely filled. Such bogs are common in New England, Wisconsin, and Minnesota.

Certain plants and animals secrete shells of carbonate of lime, and others, like single-celled diatoms, secrete shells of silica. Accumulation of such shells may appreciably aid lake-filling.

By Cutting Down Outlets. The dams of many lakes, particularly those formed of glacial débris, often consist of such loose, incoherent materials that outlet streams readily cut down into them. By this process, a lake surface may be reduced steadily until the lowest level of the lake basin is reached, causing destruction of the lake. Cutting down of outlets has been an important factor in the destruction of many

lakes, especially of glacial lakes in regions like New England, New York, Wisconsin, and Minnesota. Of course it should be borne in mind that cutting down of outlets, filling with sediment, and filling with organic remains may proceed simultaneously.

Where the dams consist of relatively hard rocks, cutting down of outlets proceeds very slowly because outlet streams contain usually very clear water whose erosive power is slight. This is true even of large rivers like Niagara and the St. Lawrence which have scarcely lowered the surfaces of the lakes they drain.

By Removal of Ice Dams. This principle is illustrated in certain regions of existing glaciers, as in the Alps, where a glacier, causing

FIG. 374. An extensive field of rough, dirty salt left in a desert basin below sea level after a lake evaporated. Bottom of Death Valley, California.

ponding of water by blockading a tributary valley, may shift position in such a manner as to allow escape of the water either under or alongside the ice.

Many ice-dam lakes, including some of great size, were either completely or largely destroyed by melting of their dams during the closing stages of the Ice Age. Thus bodies of water covering hundreds of square miles in the Black and Mohawk Valleys of New York dis-

appeared because of the removal of their great ice dams. The once vast Lake Agassiz (already described) was destroyed in a similar manner, remnants only being left (e.g., Lake Winnipeg).

By Evaporation. This is a very important method of lake destruction in arid regions where evaporation may exceed intake. Many of the depressions in the Great Basin region which once contained lakes are now dry or nearly so (e.g., Death Valley) (Fig. 374). Others now contain only small remnants of once large bodies of water, as for example the great basin of Lake Lahontan in western Nevada with its Pyramid Lake.

By Diastrophism. Ponds and small lakes are sometimes drained through fissures which are formed during earthquake disturbances. Lakes, especially larger ones, may be partly or wholly destroyed by down-warping or down-faulting of their outlet areas, but actual examples seem to be rare. It has been recently advocated that the great post-Glacial lake which once lay in the Connecticut Valley of New England disappeared by down-warping of its outlet region.

If the southern half of Florida should subside only 20 to 50 feet, most of its numerous lakes would disappear because of flooding of the region by sea water. Submergences of this kind have of course been common during geological time, but it is not easy to point definitely to recent examples of lakes thus destroyed because the evidence is so hidden. Lakes near sea level along sinking coasts are, of course, doomed to extinction.

Extinct Lakes

We have just explained the most common ways by which lakes are destroyed, and cited some examples. Among the more important criteria by which the sites of former lakes may be recognized are the following: (1) If a lake basin has been completely filled, and since then little affected by erosion, its site is marked by a flat consisting of characteristic lake deposits, practically free from boulders (Figs. 369 and 371). Such deposits may be sediments, organic (bog) materials, or salt-lake mineral deposits.

(2) Basins of larger ponds and lakes, which were not completely filled, very commonly show deposits of coarser sediments, usually in deltas and coalesced deltas, around their borders, and finer sediments, such as clays, farther out. The border deposits rise everywhere uniformly, unless subsequently affected by diastrophism, to about the former

levels of the standing waters, while the finer sediments lie at various lower levels, depending upon the topography of the lake floors.

(3) In contrast with stream deposits, lake sediments (especially the finer materials) are usually much more uniform in character and structure over wide areas.

(4) In addition to deltas, other shore features, such as wave-cut cliffs, beaches, spits, and bars, are often wonderfully preserved. This is particularly true in arid regions, as around the shores of former Lake Bonneville, Utah (Fig. 368), but they are also often well exhibited in humid regions, as around the shores of the once great Lake Agassiz.

(5) Fossils often prove that deposits were formed in lakes because many forms of life in lake waters are characteristically different from those of sea water.

CHAPTER XV

ECONOMIC GEOLOGY[1]

Introduction

In this chapter the purpose is to consider briefly geology in some of its direct relations to the arts and industries. When we realize that the value of strictly geological products taken from the earth each year in the United States alone amounts to billions of dollars, we can better appreciate the practical application of geological science. Such products include coal, petroleum, natural gas, many valuable metal-bearing minerals, and many nonmetalliferous minerals and rocks. In most cases these valuable products of nature have very slowly accumulated or concentrated at many times and under widely varying conditions throughout the millions of years of known geological time. To discover, to trace the extent of, and to remove most advantageously such deposits for the use of man are often impossible unless geological knowledge is brought to bear. Much of the practical application of geology is carried out by the mining engineer who should have, above all, a thorough knowledge of the great principles of geology.

Coal, Petroleum, and Natural Gas

Coal. Coal is the most valuable of all geological products. Although coal is, strictly speaking, not a mineral, because of both its organic origin and its lack of definite chemical composition, nevertheless it is generally classed among "mineral resources."

Coal is undoubtedly of organic (plant) origin, as shown by its composition; perfect gradations between plant deposits, like peat, and true coal; and the presence of microscopic plant remains in it. All coal represents plant material which was deposited in beds analogous to our present-day peat bogs. After a great bed of vegetable matter accumulated it was covered by sedimentary material, and thus buried

[1] Much of the material of this chapter is taken by permission from Chapter XXIII of the author's *The Story of Our Earth* which forms volume 3 of Popular Science Library published by P. F. Collier & Son Company.

in the earth's crust. Then, through very slow processes of decomposition, alteration and pressure, the vegetable matter was changed to coal. Anthracite represents the greatest degree of change in the vegetable matter.

The most perfect conditions for prolific plant growth and accumulation as great beds in the earth's crust were during the Pennsylvanian period of the late Paleozoic era in many parts of the world, but especially in the United States, China, Great Britain, and Germany. Most

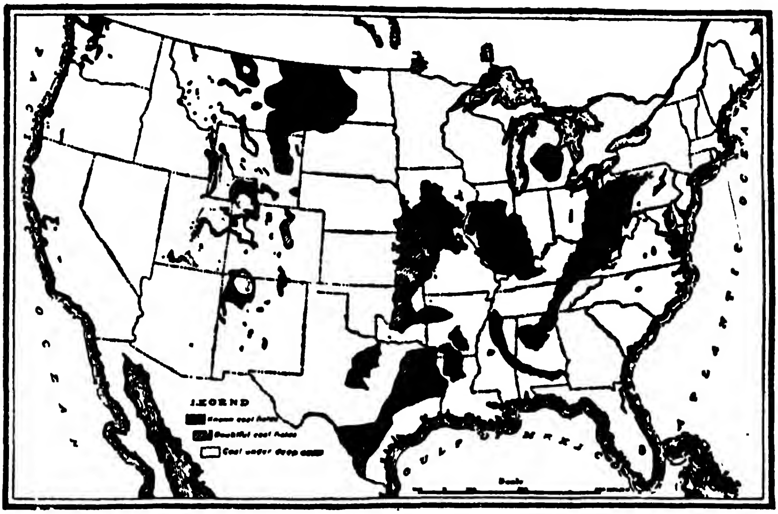

FIG. 375. Map showing the coal fields of the United States. (After U. S. Geological Survey.)

of the world's great supply of coal by far comes from rocks of Pennsylvanian age; next in importance are Cretaceous rocks; and some comes from strata of other ages later than the Pennsylvanian, even as late as the Tertiary.

The United States not only has the greatest known coal fields, but also it produces far more coal than any other country. In 1918 and again in 1944 production was nearly 700,000,000 short tons. In 1937 it was about 497,000,000 tons. Is there real danger that our supply of coal will soon run out? Hardly so, when we consider, first, the fact that probably not more than one per cent of the readily available coal has thus far been removed, and, second, the high probability that the average rate of increase in coal production for the last thirty years

will not continue. In the case of the very restricted anthracite coal fields, what might be called a crisis has already been reached because a very considerable part of the available supply has been taken out.

Approximately 350,000 square miles of the United States are underlain with one or more beds of workable coal (not including lignite)—in some areas 5 to 20 or more beds one above the other. There are also about 150,000 square miles of country underlain with the more or less imperfect coal called *lignite*. Map Fig. 375 shows the principal coal fields of the United States.

FIG. 376. An outcropping coal bed 13 feet thick, in Montana. (Photo by Campbell, U. S. Geological Survey.)

The greatest production of coal by far is from the Appalachian Mountain district, extending from Pennsylvania to Alabama, where nearly all the coal is bituminous of Pennsylvanian age. There, as well as elsewhere, the coal beds are interstratified with various kinds of sedimentary rocks, most commonly with shales and sandstones. In the Appalachian field the strata including coal beds are more or less folded toward the east, and they are nearly horizontal toward the west.

The greatest production of anthracite coal by far is from central-eastern Pennsylvania where strata of Pennsylvanian age, including a number of anthracite beds, are more or less highly folded (Fig. 377).

Less than 500 square miles are there underlain by workable anthracite coal.

Next to the greatest production of coal in the United States is from the two large areas in the middle of the Mississippi Valley (Fig. 375). It is all bituminous coal associated with nearly horizontal strata of Pennsylvanian age.

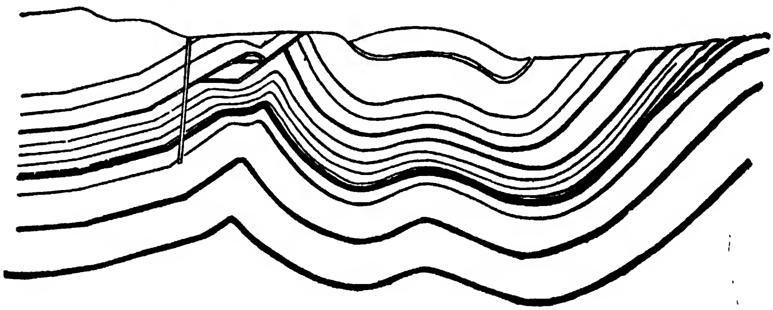

FIG. 377. Structure section more than 1000 feet deep showing folded anthracite beds of Pennsylvania. (After U. S. Geological Survey.)

The scattering areas through the Rocky Mountains yield all types of coal—anthracite, bituminous, and lignite. In some of these areas the coal beds have been but little disturbed from their original horizontal position, but usually they are more or less folded along with the en-

FIG. 378. Nearly horizontal coal beds in Indiana, with outcrops mostly concealed by glacial drift. (After G. H. Ashley.)

closing strata, the crustal disturbances affecting the coal beds having taken place late in the Mesozoic era and early in the Cenozoic era. Practically all these coals are of Cretaceous and Tertiary ages, the best being Cretaceous. Very little of the Rocky Mountain coal is anthracite.

On the Pacific Coast, coal production is relatively very small. The coals are there bituminous to lignitic of Tertiary age, usually folded in with the strata.

In Alaska there are widely distributed, relatively small, coal fields,

but they have been little developed. Alaskan coals range in age from Pennsylvanian to Tertiary, and in kind from anthracite to lignite.

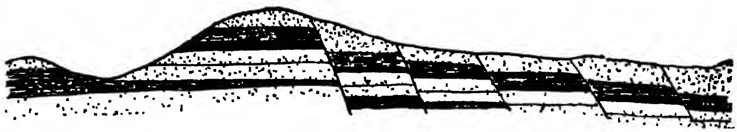

Fig. 379. Step-faulted coal beds in Scotland. (After J. C. Branner.)

Petroleum. Crude oil or petroleum is an organic substance consisting of a mixture of hydrocarbons, that is, it is made up very largely of the two chemical elements carbon and hydrogen in rather complex and variable combinations. It is practically certain that petroleum has been derived by a sort of slow process of distillation from organic matter —animal or vegetable, or both—in stratified rocks within the earth. Many strata, as for example carbonaceous shales, are more or less charged with dark-colored, decomposing, organic matter. The chemical composition itself, the kinds of rocks with which it is associated, and certain optical (microscopic) tests all point to the organic origin of petroleum. In southern California some of the oil has quite certainly been derived from very tiny oily plants, called *diatoms*, which fill many of the strata.

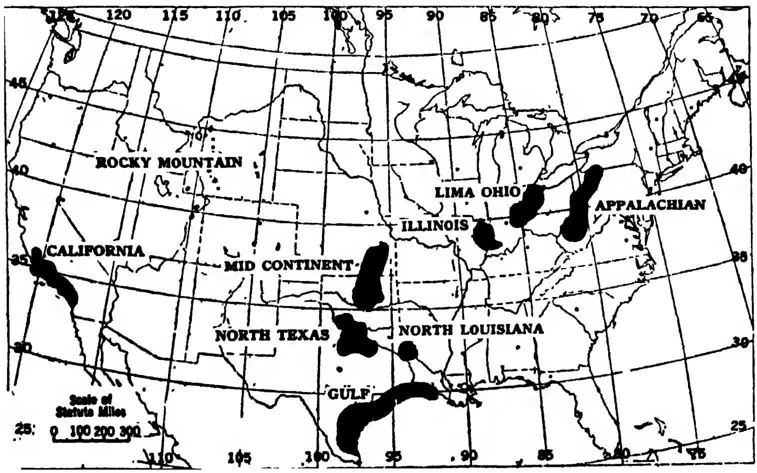

Fig. 380. Map showing the principal oil-bearing regions in the United States. (After Geological Survey of Kansas.)

During the last forty years petroleum has come to be one of the most important and useful natural products. Among the many substances artificially derived from petroleum are kerosene, gasoline, naphtha, benzine, vaseline, and paraffine. The United States greatly leads in the production of petroleum, but various other countries are also important producers. Map Fig. 380 shows the regions of the United States which contain the principal oil fields. The total area of the fields underlain with oil is several thousand square miles.

In the Appalachian, Ohio-Indiana, Illinois, Mid Continent, and North Texas fields the strata carrying oil range in age from Ordovician to Pennsylvanian, and they are, for the most part, only moderately disturbed from their original position. The oil from the Gulf region comes mainly from Cretaceous and Tertiary strata with a gentle downtilt under the Coastal Plain toward the Gulf of Mexico. In California the oil-bearing strata are of Tertiary age, and they have been generally much disturbed and folded.

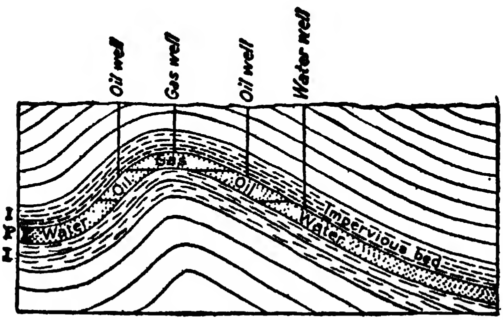

FIG. 381. Structure section illustrating a very common mode of occurrence of oil. (After U. S. Geological Survey.)

Under proper conditions below the earth's surface, the derived oil accumulates in porous or fractured rocks. There must, of course, be a source from which the petroleum is derived or distilled; a porous or fractured rock formation to take it up; a cap-rock or impervious layer to hold it in; and a proper geologic structure to favor accumulation. The most common porous (containing) rock is sandstone, and the most common cap-rock is shale. "Oil is rarely found without gas, and saline water is likewise often present. If the containing strata are horizontal, the oil and gas are usually irregularly scattered, but if tilted or folded, and the beds porous throughout, they appear to collect at the highest point possible. It was the result of observations along this line that led I. C. White to develop what is known as the 'anticlinical theory.' According to this theory, in folded areas the gas collects at the summit of the fold (anticline), with the oil immediately below, on either side, followed by the water (Fig. 381). It is, of course, necessary that the oil-

bearing stratum shall be capped by a practically impervious one. If the rocks are dry, then the chief points of accumulation of the oil will be at or near the bottom of the syncline (downfold) or lowest portion of the porous bed. If the rocks are partially saturated with water, then the oil accumulates at the upper level of saturation. In a tilted bed, which is locally porous (Fig. 382), and not so throughout, the oil, gas, and water may arrange themselves according to their gravity in this porous part" (H. Ries).

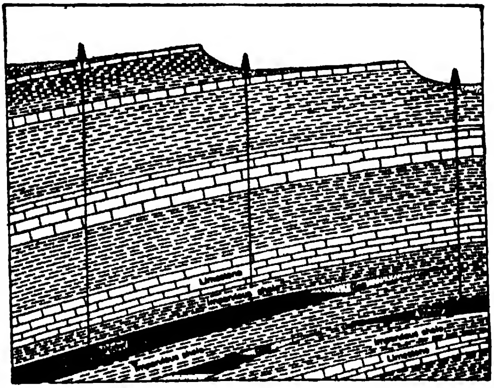

FIG. 382. Structure section illustrating a common mode of occurrence of oil. (After Geological Survey of Kansas.)

Although the term "oil-pool" is commonly used, there is really no actual pool or underground lake of oil, but rather there is porous rock saturated with oil. Sometimes the oil is under great pressure (either gas or hydrostatic pressure) and the oil shoots into the air, forming a gusher (Fig. 383), when a well taps an "oil-pool." It has been estimated that, in an oil field of average productiveness, a cubic foot of the porous rock contains from 6 to 12 pints of oil. The life of a well drilled into an "oil-pool" varies from a few months to 20 or 30 years, or sometimes even more. A heavy producer (especially a gusher) almost invariably falls off in production notably in a few months or at most in a few years. Producing wells range in depth to fully two and one-half miles. In 1944 the United States produced 1,677,000,000 barrels of oil of 42

FIG. 383. An oil gusher in the Sunset-Midway district, California, in 1910. (Photo by R. W. Pack, U. S. Geological Survey.)

gallons each, or six times as much as Russia, the second country in rank. In 1944 the six leading producing states, in order of importance, were Texas, California, Louisiana, Oklahoma, Kansas, and Illinois. Other leading producing regions in 1944, in order of importance, were Venezuela (73,000,000 barrels), Iran, East Indies, Mexico, Rumania, Iraq, Argentina, Trinidad, and Colombia (22,000,000 barrels).

Natural Gas. The most perfect fuel with which nature has provided us is natural gas. Not only is it easily transported even long distances through pipes, but also as a fuel it is easily regulated, leaves no refuse, and is less damaging to boilers than coal. It is a colorless, odorless, free-burning gas consisting very largely of the simple hydrocarbon called marsh gas or fire damp. Petroleum nearly always has more or less natural gas associated with it, but in some places considerable quantities of gas may exist alone. Natural gas like petroleum is of organic origin—a product of slow natural distillation of vegetable or animal matter, or both, within the earth's crust.

One of the most common modes of occurrence of gas is at the top of an anticline (up-fold) in porous rock (like sandstone) between impervious layers (like shale). Fig. 381 well illustrates the principle, the gas lying above the oil, and the oil above the water, that is, the three substances are arranged according to specific gravity. Gas may also exist in considerable quantities in irregular bodies of porous or fractured rocks. Natural gas is nearly always under pressure within the earth, hundreds of pounds per square inch being common, while more exceptionally, as in certain West Virginia wells, pressures of over 1000 pounds have been registered.

The United States is by far the greatest world producer of natural gas, the output for 1943 having been 5,000,000,000,000 cubic feet. Texas, Oklahoma, California, and Louisiana were the greatest producers. Areas underlain with natural gas are, in the main, the same as for petroleum, and they total several thousand square miles.

Metal-bearing (Ore) Deposits

Iron. Without question, the most useful of all metals is iron. As such, it is rare in nature, but in chemical combination with other substances it is extremely widespread and very common. Iron makes up about 5 per cent of the weight of the earth's crust, but in the form of *ore* (i.e., a metal-bearing mineral or rock of sufficient value to be

mined) it is notably restricted in occurrence. The three great ores of iron are the minerals hematite, magnetite, and limonite whose composition and characteristic properties are given in Appendix A.

The United States is by far the greatest producer of iron ore in the world, the output for 1942 having been about 100,000,000 long tons, the greatest in the history of this or any other country. In 1928 the production was 62,000,000 tons, and in 1937, 73,000,000 tons. Most of it was hematite ore. Other important producers of iron ore are France, Russia, Sweden, Great Britain, and Germany.

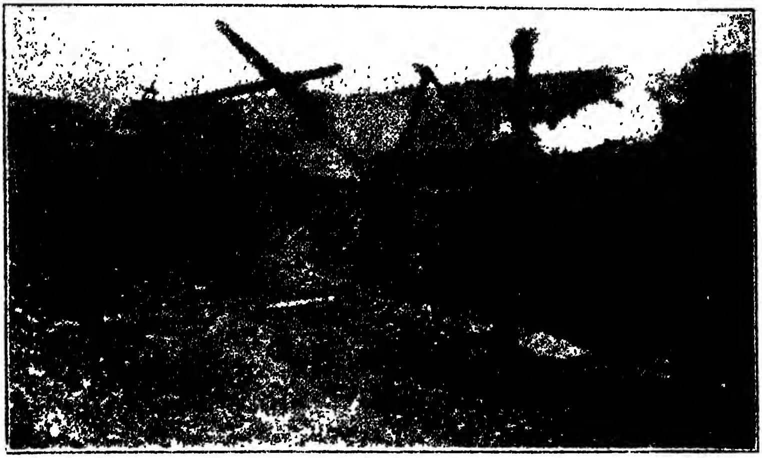

FIG. 384. Steam shovels removing soft, rich, hematite iron-ore in the Mesabi district, Minnesota. (Photo by F. C. Adams.)

We shall now very briefly consider the several chief iron-mining districts of the United States, giving some idea of the modes of occurrence and origin of the ores. Greatest of all is the Lake Superior region not far west and south of the lake in Minnesota, Michigan, and Wisconsin. Considerably more than half of the iron mined in the United States comes from the single state of Minnesota, and about one-fourth of it from Michigan. The Minnesota deposits are of irregular shape lying at or near the surface (usually covered only by glacial deposits). Few of them extend downward more than some hundreds of feet. The soft high grade ore is removed by steam shovels in great open pits (Fig. 384). In the several districts of northern Michigan and Wisconsin, the ores (nearly all hematite) are associated with more or less highly folded rocks at considerable depths. The Lake Superior

iron ores all occur in rocks of Archeozoic and Proterozoic ages. According to the best explanation of their origin, the iron of the ores was once part of a sedimentary series of rocks in the form of iron carbonate and silicate interstratified with layers of a flintlike rock, and associated with slate, quartzite, etc. After these rocks were raised into land and subjected to weathering, the old iron compounds were altered to oxides, mainly hematite, and somewhat concentrated. Further concentration of the ore was caused by dissolving out the flintlike layers of the old rocks (Fig. 385).

The Birmingham, Alabama, region is the second most important iron ore producer in the United States. The ore is hematite, forming part of a Silurian formation. The ore appears to be an original bed (or locally several beds) of fairly rich iron ore which was deposited on the shallow Silurian sea bottom and then covered by other strata. At the close of the Paleozoic era the iron ore was more or less highly folded in with other strata throughout the Appalachians. A remarkable fact regarding the Birmingham district is that, in the near vicinity of the ore, there are both coal for fuel and limestone flux for smelting the ores.

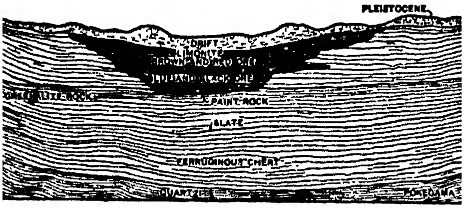

FIG. 385. Structure section showing the mode of occurrence of hematite ore at Mesabi, Minnesota. (After Leith, U. S. Geological Survey.)

Another fairly important iron-mining district of the United States is the Adirondack Mountain region of northern New York. Magnetite is the ore (Fig. 386), and it occurs in more or less irregular lenses and bands in granite and other closely associated rocks of pre-Paleozoic age. One view regarding the origin of this ore is that it segregated during the process of cooling of the molten granite, and another view (advocated by the author) is that it was derived from an older iron-rich igneous formation either by the molten granite or by very hot solutions from it and concentrated into the ores.

The third important iron ore is limonite. Most of it, in the United States, comes from the Appalachian region. It is all of secondary origin, that is, it has been derived from certain early Paleozoic iron-bearing limestones either by weathering or solution and concentrated into ore deposits.

Copper. This is one of the most useful of all metals. Various minerals containing copper are found in many parts of the world, but only about six of them are really important as ores. These are native copper, chalcopyrite, chalcocite, azurite, malachite, and cuprite, most of which are described in Appendix A. The number of places where they may be profitably mined as ore is distinctly limited. Fifteen or twenty countries produce more or less copper, but the United States is by far the greatest producer. The output was about 2,000,000,000 pounds in 1944 and 1,684,000,000 pounds in 1937. Other leading producers are Chile, Rhodesia, Canada, Belgian Congo, Russia, and Japan. The principal producing states are Arizona, Utah, Montana, Nevada, New Mexico, and Michigan.

FIG. 386. A specimen of magnetic iron-ore from Lyon Mountain, New York. Most of the black material is magnetite.

In Arizona several great copper mining districts lie in the southeastern one-fourth of the state. Almost invariably the ores are directly associated with limestone and an igneous rock (granite), both of late Paleozoic age (Fig. 387). The ores are almost always near the border between the two rocks, mostly as great irregular deposits within the limestone, and less commonly as veins within the granite. The original ores were carried in solution and deposited by hot liquids (or vapors) from the cooling granite.

In the region around Butte, Montana, most of the ores are sulphides of copper (mainly chalcocite) which occur with quartz in a great system of nearly parallel veins in granite of Tertiary age. "It is supposed that in the copper veins the hot ore-bearing solutions ascended the fractures

in the (hot) granite, replacing the rock by ore, and resulting in an intense alteration of the walls."

In Michigan the mines are located on Keweenaw peninsula which extends into Lake Superior. A unique feature is that the ore is native copper associated with some native silver. The rocks containing the ore are steeply tilted lava sheets and conglomerate strata of Proterozoic age. Openings in porous lava and spaces between the conglomerate pebbles have been filled by metallic copper which was carried off in hot solutions from the cooling lavas. Certain of the mining shafts have been sunk more than 5000 feet below the surface, these being among the deepest in the world.

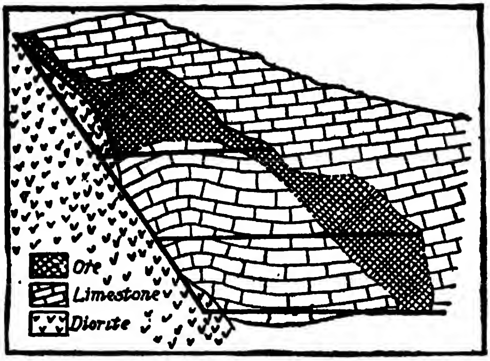

FIG. 387. Structure section showing the mode of occurrence of copper ore at Globe, Arizona. (After Wendt.)

In Utah the greatest mining district is at Bingham Canyon, southwest of Salt Lake City. The rocks are late Paleozoic strata pierced by a large body of igneous rock. Some of the sulphide ores (mainly chalcopyrite) occur in veins in the igneous rock, and some in large tabular masses in the adjacent limestone. Hot solutions from lower portions of the uncooled igneous rock carried the ore in solution into the limestone and also into cracks in the upper cooled igneous rock.

Lead. Lead must surely be counted among the five or six most useful metals. As in the case of nearly all the other most important natural resources, the United States is the world's greatest producer of lead. Its output of lead was 655,000 short tons in 1925, and 417,000 tons in 1944. Most of it came from Missouri, Idaho, Utah, and Oklahoma. Leading foreign countries are Canada, Belgium, Germany, Mexico and Australia. Nearly all the lead comes from the mineral galena (a sulphide of lead) which is described in Appendix A.

In the great Joplin region, including parts of Missouri, Oklahoma, and Kansas, the ore (galena), associated with much zinc ore, occurs as

veins and great irregular deposits in limestone of early Paleozoic age. It is generally thought that underground waters dissolved the ores out of the limestone in which they were disseminated as tiny particles and deposited them in concentrated form at lower levels.

In the famous Coeur d'Alene district of northern Idaho the great output of lead is obtained from a lead-silver ore, that is galena rich in silver. It occurs in great fissure veins mostly following fault fractures in highly folded strata of Proterozoic age. Igneous rocks cut through the strata, and it is believed that hot ore-bearing solutions given off from the highly heated igneous rocks rose in the fissures and deposited the ores.

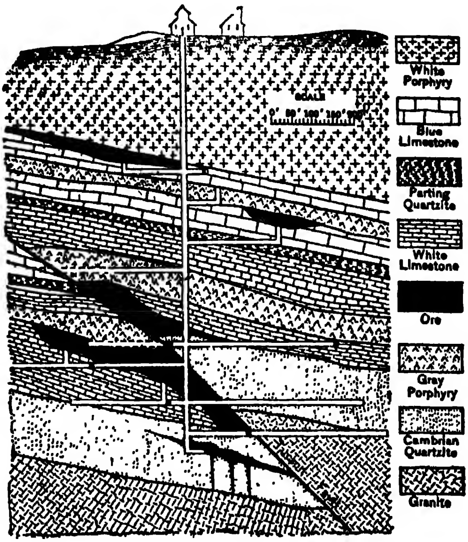

FIG. 388. Structure section showing the mode of occurrence of lead and zinc ores at the Tucson Shaft, Leadville, Colorado. (After Argall.)

The Park City and Tintic districts of Utah are great producers of lead. The lead ore (galena) is usually rich in silver. It occurs mainly in veins and irregular deposits in limestone of Paleozoic age closely associated with certain igneous rocks.

One of the most famous mining districts in the world is that around Leadville, Colorado, where ores of four metals—gold, silver, lead, and zinc—have been extensively mined. The salient points in the rather complex geology are the following: Paleozoic strata, including much limestone, rest upon a foundation of pre-Paleozoic granite. Sheets of igneous rock are interbedded with the strata, and many dikes of igneous rocks cut through the whole combination. After the last igneous activity, all the rocks were somewhat folded and notably faulted in many places. The ores were dissolved out of the igneous rocks and deposited in large masses mostly in the limestone and in fissure veins especially along and near the fault zones (Fig. 388).

Zinc. Another of the few most useful metals is zinc. It never occurs in metallic form in nature, but most of it by far is obtained from the ore-mineral sphalerite, described in Appendix A. A red oxide of zinc ore, called *zincite,* assumes great economic importance in New Jersey.

In 1944 the United States produced 719,000 short tons of zinc, 630,000 tons in 1929, and 626,000 in 1937. The United States is the world's greatest producer. Leading states are Missouri, Oklahoma, New Jersey, Idaho, and Tennessee. Belgium, Poland, Canada, and Germany are great foreign producers.

Most important of all in the United States in recent years is the Joplin region (Tri-state), including parts of Missouri, Oklahoma, and Kansas, where the ore is closely associated with lead ore. The mode of occurrence and origin of these ores are above refered to in the discussion of lead.

In the Butte, Montana, region, some of the great east-west fissure veins in granite are rich in silver ores in the upper levels, and in zinc ores (mainly sphalerite) at depths of from some hundreds of feet to nearly 2000 feet, that is, as far down as they have been mined. They, like the great copper veins of the same general district, were carried by hot solutions which rose from the lower still very hot granite and deposited the ores in fissures of the same cooler rock higher up.

In the general vicinity of Franklin, New Jersey, the zincite ore-deposits occur in white limestone along, or close to its contact with, metamorphosed (altered) strata and granite of early Paleozoic age. It is not definitely known how the ore originated, but it was probably derived in solution from the hot granite, and deposited in the limestone by replacement of the latter.

In Colorado the principal zinc mines are around Leadville where lead ore is nearly always directly associated with the zinc ore. This district is above described in the discussion of lead.

Gold. This precious metal has been used and highly prized by man for thousands of years. The Transvaal region of South Africa has for many years been the world's greatest gold producer. In 1937 South Africa produced gold to the value of $400,000,000; Canada, $140,000,000; and the United States, $170,000,000. Russia, Mexico, and Australia are also important producers. The leading producing states are California, South Dakota, Colorado, Arizona, and Nevada. Alaska is also an important producer. During World War II, gold

production fell off greatly. Since the war, particularly in 1946 and 1947, the most spectacular gold-rush and development has been on the doorstep of the Arctic Circle in the Northwest Territories of Canada, centered at Yellowknife on the north shore of Great Slave Lake.

Most of the commercially valuable gold occurs in nature as native gold either mixed with gravel and sand (i.e. placer deposits) along existing or ancient stream beds, or in veins mechanically held in the mineral pyrite (described in Appendix A) in submicroscopic form, or mixed with quartz in vein deposits. In deep vein deposits it is quite the rule to find free (or native) gold mechanically and visibly mixed with quartz in the upper levels, while deeper down the gold is mechanically, but invisibly, held in combination usually in pyrite, which latter is associated with quartz. This difference is due to the fact that the lower-level ores are now just as they were formed, but in the upper levels the ores have been weathered and the gold set free and often more or less further concentrated by solutions. Vein deposits are found in many kinds of rocks—igneous, sedimentary, and metamorphic—of nearly all ages, though they are generally directly associated with igneous rocks. In nearly all cases the best evidence indicates that the vein fillings were formed by hot ore-bearing solutions from the igneous rocks which deposited the ore plus quartz in fissures in either the igneous rocks or adjacent rocks. Among the many localities where fissure veins of the kind just described are of great economic importance are the "Mother Lode" belt of the Sierra Nevada Mountains of California; Cripple Creek, Georgetown, and the San Juan region of Colorado; and Goldfield, Tonopah (Fig. 389), and Comstock Lode of Nevada.

Placer deposits, that is, free gold mixed with gravel and sand, also yield much gold. They are most prominently developed in California and Alaska. These gold-bearing "gravels represent the more resistant products of weathering, such as quartz and native gold, which have been washed down from the hills on whose slopes the gold-bearing quartz veins outcrop and were too heavy to be carried any distance, unless the grade was steep. They have consequently settled down in the stream channels, the gold, on account of its higher specific gravity, collecting usually in the lower part of the gravel (placer) deposit" (Ries). Such gold occurs as grains, flakes, or nuggets.

Most of the gold of South Africa comes from the Witwatersrand district where the native metal occurs in a unique manner in beds or layers of conglomerate associated with other strata, all the rocks being considerably folded and somewhat faulted. Some of the mines are sev-

eral thousand feet deep. The gold either accumulated in placer form with gravel which later consolidated into conglomerate or it was introduced into spaces between the pebbles subsequently by ore-bearing solutions.

Fig. 389. Gold-bearing quartz veins in Bullfrog Mine, Rhyolite, Nevada. (After F. L. Ransome, U. S. Geological Survey.)

Silver. For many years Mexico and the United States have been the world's greatest silver producers. In 1937 silver production in Mexico was 85,000,000 troy ounces, and in the United States 71,000,000 troy ounces. In 1944 their production was 65,000,000 and 36,000,000 troy ounces, respectively. Canada and Australasia are also important producers. Important producing states are Idaho, Utah, Arizona, Montana, Colorado, and Nevada.

In Montana most of the silver is in the native form, more especially in the upper portions of the great veins rich in copper and zinc ores near Butte. These ores and their origin are described above under "Copper" and "Zinc."

The two greatest silver districts of Nevada are Tonopah and Comstock Lode where silver and gold minerals are associated as ores in Tertiary igneous rocks, the ores having been deposited in veins by hot ore-bearing solutions from the igneous rocks.

In Idaho the Coeur d'Alene district produces most of the silver, the ore there being a silver-bearing lead ore (galena). The nature and origin of these deposits are described above under "Lead."

In Utah the silver is also obtained from silver-bearing galena especially in the Tintic, Cottonwood Canyon, and Bingham Canyon districts where the ores occur mainly as irregular deposits and in fissure veins in Paleozoic strata (chiefly limestone) directly associated with igneous rocks, hot ore-bearing solutions from the igneous rocks having furnished the ores.

Tin. Production of tin in the United States has never amounted to much, a little mining having been carried on from time to time in South Carolina, Black Hills of South Dakota, and southern California. The Malay region, Bolivia, and the East Indies are the principal producers of tin. The figures in long tons for 1937 were Bolivia, 25,000; Malay region, 77,000 and East Indies, 39,000. During World War II production in Bolivia increased notably, but it fell off greatly in Malay and the East Indies.

The only important ore of tin is the mineral cassiterite, described in Appendix A. In the Malay region the ore all occurs in placer deposits and is, therefore, of secondary origin, the source of the ore not being known. In Bolivia the tin ore occurs in veins in, and close to, granite, the ore having been carried by very hot vapors or liquids which were derived from the still highly heated granite.

Aluminium. The mineral called bauxite (a hydrous oxide of aluminum) is the great ore from which aluminum is obtained by an electrical process. Bauxite is noncrystalline, relatively light in weight, white to yellowish in color, and in the form of rounded grains or earthy or claylike masses. The United States, France, Dutch Guiana, Italy, Yugoslavia, and Hungary are the greatest producers. In the United States the principal deposits are in Arkansas, Georgia, and Alabama. Bauxite is probably always a secondary mineral formed by decomposition of igneous rocks rich in certain aluminum silicate minerals. In some cases, as in the Georgia-Alabama region, the bauxite appears to have been formed and concentrated in deposits by hot solutions from uncooled igneous rocks. In 1944 the United States produced 3,200,000 tons of aluminum ore.

Mercury. This metal, commonly known as "quicksilver," is of special interest because it is the only one which exists in liquid form at ordinary temperatures. The metal occurs in only small quantities in

nature, most of it by far being obtained from the red mineral called cinnabar, described in Appendix A.

The greatest quicksilver producing countries are Italy, Spain and the United States. In the United States, California is by far the leading state, while Oregon and Texas are the only other important producers.

In California most of the ore occurs in veins and irregular deposits in metamorphosed strata of Mesozoic and Cenozoic ages usually closely associated with igneous rocks. There, as well as in other parts of the world, hot vapors from igneous rocks carried the volatile ore upward and deposited it in fissures.

Uranium. World War II saw the significance of uranium in atomic research and development. America's great known source of this rare element is in a complex ore body located in a remote district at the east end of Great Bear Lake, on the Arctic Circle, northwest Territories, of Canada. The ore there contains the mineral pitchblende which is the oxide of uranium.

Other Economic Products

Building Stones. Some of the principal features which should be considered in regard to building stones are power to resist weathering; power to withstand heat; color; hardness and density; and crushing strength. Building stones representing rocks of nearly all important geologic ages are widely distributed throughout the world.

Granite, including certain other closely related rocks, is one of the oldest and most useful building stones. The New England states are the greatest producers, while the Piedmont Plateau district (east of the Appalachians) from Philadelphia to Alabama also contains important granite quarries. In the Adirondack Mountains, in Wisconsin and Minnesota, through the Rocky Mountains, and in the Sierra Nevada Mountains there are extensive areas of granite which are relatively little quarried. The granite usually occurs in regions of highly disturbed rocks where great volumes of the molten material were forced into the earth's crust, cooled, and later laid bare by erosion.

Marble, according to the geological definition, is a metamorphosed limestone, that is a crystalline limestone. More loosely in trade, any limestone which takes a polish may be called marble. The greatest marble-producing districts of the United States are western New England (especially Vermont), the Piedmont Plateau, and Appalachian Mountains, all in rocks of Paleozoic or greater age. In northern New

York and in the mountains of the west there are relatively few marble quarries.

Ordinary *limestones* are widely distributed in many states where they range in age from early Paleozoic to Tertiary. Most of the quarries supply stone for near-by markets. The so-called Bedford limestone of Indiana has, for many years, been a widely used limestone for building purposes in the United States and many other parts of the world (Fig. 390).

FIG. 390. A limestone quarry for building stone. Bedford, Indiana. (Photo by courtesy of the Bedford Quarries Company.)

Sandstones, which are stratified rocks consisting mainly of rounded quartz grains cemented together, are widely used in building operations. Like limestones, they are very widespread in formations of all ages except the very oldest. There are many standstone quarries supplying more or less local markets throughout the country. Two of the best known and most widely used sandstones are the so-called brownstone of Triassic age extending interruptedly from the Connecticut Valley of Massachusetts to North Carolina, and the Berea, Ohio, sandstone of light gray color and uniform texture.

Slate is mostly a metamorphosed shale, that is, a shale which has been subjected to great pressure within the earth so that the stratification has often been obliterated, and a well defined cleavage has been

developed at right angles to the direction of application of the pressure. Good slate is fine-grained, dense, and splits readily into wide, thin plates. It occurs only where mountain-making pressure and metamorphism have been brought to bear upon the strata. Most of our great slate quarries are located in early Paleozoic rocks of New England (Fig. 391), eastern New York, and southward through the Piedmont Plateau. Some quarries are also located in Arkansas, Minnesota, and California.

Clay. Most clays originate by the weathering of rocks, particularly igneous and metamorphic rocks rich in the mineral feldspar. As a result

FIG. 391. A roofing slate quarry. Brownsville, Maine. (After T. N. Dale, U. S. Geological Survey.)

of the decomposition of the feldspar, much clay is formed, the main substance of which is kaolin. Both feldspar and kaolin are described in Appendix A. When the resulting clay rests upon the rock from which it has been derived it is called residual clay. Much of the clay is, however, carried away, mainly by streams, and deposited in lakes or the sea or on river flood-plains. Some clay deposits are of wind-blown origin, and still others are formed by the grinding action of glaciers. Clays are very widespread, and they are directly associated with rocks of all geologic ages.

Lime and Cement. Limestone, which is one of the most common and widespread of all stratified rocks, forms the basis for the manufacture of the important substances *lime* (or "quicklime") and *Portland cement.* Lime results when pure limestone (carbonate of lime) is "burned" or heated to a temperature high enough to drive off the carbonic acid gas.

Certain limestones containing clay of the right kind and proportion are called *natural cement rocks* because, after being "burned," they develop the property of "setting" like cement when mixed with water. The "setting" of a cement is due to the fact that certain chemical compounds, formed during the heating, crystallize when mixed with water, and the hard, tiny interlocking crystals of the newly formed silicate minerals give rigidity to the mass. Of recent years Portland cement has largely superseded the natural rock cements. "Portland cement is the product obtained by burning a finely ground artificial mixture consisting essentially of lime, silica, alumina, and some iron oxide, these substances being present in certain definite proportions" (Ries). The necessary ingredients are generally obtained by grinding and burning carefully selected mixtures of limestone in some form with clay or shale.

Salt. Most of the common salt (the mineral *halite*) of commercial value occurs in nature in sea or salt-lake water; or in beds of *rock salt* associated with other strata; or as *natural brine* in openings or pores in certain rocks. Considerable salt is obtained by evaporation of tide water, as around San Francisco Bay, and of salt-lake water, as at Great Salt Lake, Utah. The salt of a salt lake has been washed out of the rocks of the surrounding country and gradually accumulated in the lake because it has no outlet.

Most important of all sources of salt is the rock salt which occurs in the form of strata within the earth's crust. Such strata are found in rocks of nearly all ages from the early Paleozoic to the present. They have resulted from the evaporation of salt lakes or salty, more or less cut-off, arms of the sea, after which other strata have accumulated on top of them. Thus in the Silurian system of nearly horizontal strata underlying all of southwestern New York state there occur almost universally from one to seven beds of salt. At Ithaca, New York, seven salt beds were struck in a well at a depth of about 2200 feet. One well in central-western New York penetrated a layer of solid salt 325 feet thick. Some of this New York salt is being mined much

like coal, but most of it is obtained by running water into deep wells to dissolve the salt, the resulting brine being pumped out and evaporated.

Under portions of southern Michigan, salt occurs both in beds and in natural brines charging certain porous rock layers. Both the salt beds (of Silurian age) and the brines (of Mississippian age) supply great quantities of salt. The brines are pumped out and evaporated.

In 1944 the United States produced 16,000,000 short tons of salt. Michigan, New York, and Ohio are the leading producers, but many other states are important producers.

Gypsum. The composition and properties of this common and useful mineral are given in Appendix A. *Rock gypsum* is the variety of great commercial importance. It is widespread, being quarried in many states, and occurs interstratified with rocks of many ages where it has originated by evaporation or partial evaporation of salt-water lakes or more or less cut-off arms of the sea. Salt beds are often associated with gypsum.

For about thirty years (including 1944) the average yearly production of gypsum in the United States has been several million long tons, or about eight times that of Canada, the nearest competitor. New York, Iowa, Michigan, and Ohio are the chief producers. In New York the rock gypsum (usually 4 to 10 feet thick) lies in layers between shale and limestone strata of Silurian age, and it is quarried from the central to the western part of the state. In Michigan the rock-gypsum beds, commonly 5 to 20 feet thick, lie in Mississippian strata in the southern portion of the state. A great bed of exceptionally pure rock gypsum underlies about 25 square miles of Webster County, Iowa, in strata of late Paleozoic age. The Kansas gypsum deposits extend across the central part of the state in rocks of Permian age.

Rock gypsum is mainly used in making "plaster of Paris," as a retarder in cement, and as a fertilizer (so-called "land plaster").

APPENDIX A

Some Common and Useful Minerals

A reasonable acquaintance with the more common and useful minerals can be gained only by a study of actual specimens. In the following list, alphabetically arranged for convenience of reference, only the more obvious, easily determinable properties of each mineral are listed.

Amphiboles. A number of species related in composition, crystal form, and properties are here included. They are mostly very complicated silicates of calcium and magnesium usually also with aluminum, sodium or iron. The most common and important one is called *hornblende.* It crystallizes with well-defined prismatic faces (Fig. 8a), and with two good cleavages crossing at angles of about 124° and 56° and parallel to the prismatic faces. Color, dark brown to black. Transparent to opaque. Hardness, nearly 6. Specific gravity, over 3. It is a very common mineral, especially in igneous and metamorphic rocks. Much less common varieties of amphibole are *tremolite,* colorless to light gray, and common in metamorphic limestones; actinolite, a green variety common in some metamorphic rocks; one kind of *asbestos* which is a very fibrous mineral; and one kind of *jade* which is white, gray or green and very tough.

Apatite. Composition: a combination of phosphorus, oxygen and calcium. $Ca_4(CaF)(PO_4)_3$. Crystallizes in regular six-sided prisms capped at each end by a six-sided pryramid (Fig. 392). Color variable, but usually green or brown. Transparent to opaque. Hardness, 5. No good cleavage. Specific gravity, 3.2. Rather widely disseminated in many common kinds of rocks. Apatite, mainly in uncrystallized form, is the chief source of phosphate fertilizer.

Azurite. Composition: a compound of copper, hydrogen, carbon, and oxygen. $2CuCO_3.Cu(OH)_2$. Commonly crystallized in tabular and prismatic forms. Color, characteristic azure-blue. Translucent to opaque. Hardness, nearly 4. Specific gravity, nearly 4. No good cleavage. Commonly occurs in veins. It is an ore of copper, as for example in some of the copper mines of Arizona.

Barite. A compound of barium, sulphur, and oxygen crystallizing in tabular prismatic forms. $BaSO_4$. Often called "heavy spar" because of its specific gravity of 4.5 which is notably greater than that of the average light-colored mineral. It has three good cleavages, two of them at right angles. White or colorless when pure. Transparent to opaque. Hardness, 3.5. It is a common mineral, especially in vein deposits, often associated

with ores. Used in powdered form to give added weight to certain kinds of paper and cloth. It is the source of a barium compound used to refine sugar.

Beryl. Composition: a complex combination of silicon, oxygen, aluminum, and the rare element beryllium. $Be_3Al_2Si_6O_{18}$. Usually crystallizes in regular six-sided prisms, sometimes a foot or more long (Fig. 8c). Color, usually white, green, blue or yellow. Transparent to translucent.

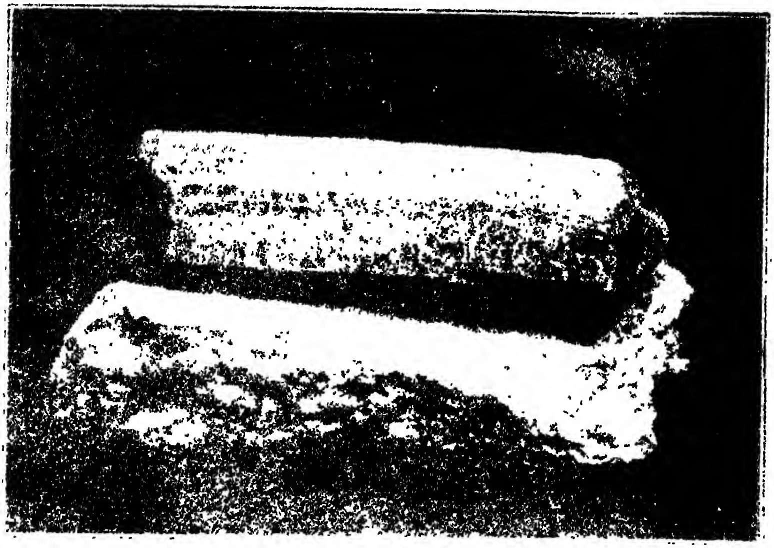

FIG. 392. A crystal of apatite on limestone. (Courtesy of Katharine Bryant.)

Specific gravity, 2.8. Practically no cleavage. Very exceptionally hard, being 8 in the scale. Two varieties of beryl—*emerald* (green) and *aquamarine* (blue)—are well-known and highly prized gem stones. The colors are due to slight impurities. Beryl occurs most commonly in a special kind of coarse grained granite, and in metamorphic rocks.

Calcite. Sometimes called "calc spar." Composition: a combination of calcium, carbon, and oxygen. $CaCO_3$. Exhibits a great variety of crystal forms, but all of them with crystal faces arranged in sixes or threes around a principal axis. Figures 8e,f,g illustrate a few of the most common shapes. Three almost perfect cleavage directions, none meeting at right angles. Color, white or colorless when pure, but of various shades when impure. Transparent to opaque. Hardness, 3. Specific gravity, 2.7. Bubbles freely when touched with a drop of cold, dilute, hydrochloric acid. Calcite is a very common mineral especially because much limestone (in-

cluding *chalk*) and marble usually consist largely of it. Very common as vein fillings (Fig. 333). Also often occurs in the form of a stringy, porous spring deposit called *travertine*, and as cave deposits such as *stalactites* and *stalagmites* (Fig. 332). A very clear, crystalline variety is *Iceland spar*. Calcite is a very useful mineral widely employed in the form of limestone and marble for building stone, decorative purposes, etc. Also used in making quicklime, as a flux in smelting certain ores, in glass making, etc.

FIG. 393. A group of calcite crystals. (Courtesy of the American Museum of Natural History.)

Cassiterite. Composition: a simple compound of tin and oxygen. SnO_2. Crystallizes in four-sided prisms capped with pyramids. Hardness, over 6 (greater than steel). Specific gravity, 7 (notably high). Color, brown to nearly black. Translucent to opaque. Practically no cleavage. Usually occurs in granite or metamorphic rocks near granite, and also in gravel deposits, as in the important mines in the Malay region. It is the one great ore of tin.

Chalcocite. Composition: a simple compound of copper and sulphur. Cu_2S. Crystallizes in flattish prismatic forms, but usually not crystallized. Shiny black with metallic lustre, and tarnishes on exposure to air to dull black. Opaque. Hardness, nearly 3. Specific gravity, nearly 6 (notably high). No cleavage. Occurs in vein deposits often as an important ore of copper, as at Butte, Montana.

Chalcopyrite. Sometimes called "copper pyrites." Composition: a simple compound of copper, iron, and sulphur. $CuFeS_2$. Crystallizes in four-sided tetrahedron-like forms, but good crystals are not common. Color, characteristic deep brass-yellow with metallic lustre. Opaque. Hardness, 3.5. Specific gravity, over 4. Rather widely distributed, usually in vein deposits, often associated with other metal-bearing minerals. It is a very important ore of copper, as at Rio Tinto, Spain.

Cinnabar. Composition: mercury and sulphur compound. HgS. Color, characteristic vermilion-red. Transparent to opaque. Hardness, 2.5, being an extra-soft metal-bearing mineral. Specific gravity, over 8, being extra-heavy. Small three-sided crystals are rare. Completely vaporizes on being heated. It is the one great ore of mercury (quicksilver), especially in California and Spain.

Copper. Known as "native copper." Composition: copper only. Color, characteristic copper-red, with metallic lustre. Opaque. Hardness, less than 3. Specific gravity, nearly 9, being extra-heavy. Cubic and modified cubic crystals are not common. Rather widely distributed, usually in veins. It is an important ore of copper, especially in the great mines of northern Michigan.

Corundum. Composition: a combination of aluminum and oxygen. Al_2O_3. Crystallizes in six-sided prisms capped with very steep pyramidal faces (Fig. 8d). Hardness, 9, being among the few very hardest of all known minerals. Specific gravity, about 4. Three good cleavages make angles of nearly 90° with each other. Color, usually gray to brown, but varies greatly. Transparent to opaque. Two of the most highly prized precious stones—*ruby* (red) and *sapphire* (blue)—are nearly transparent, slightly impure varieties of corundum. *Oriental topaz* (yellow), *oriental emerald* (green), and *oriental amethyst* (purple) are also varieties of corundum. *Emery* is a fine grained, gray, crystalline variety of corundum usually mixed with magnetite, etc. Corundum occurs in various igneous and metamorphic rocks, and also in gravel deposits, as in the sapphire and ruby mines of Burma, Siam, and Ceylon. Emery was formerly mined in Asia Minor, Massachusetts, and California for use in the manufacture of abrasives.

Diamond. Composition: pure carbon. Crystallizes in octahedral forms. Colorless when pure, but often variously tinted, one variety being almost black. Transparent to opaque. Exceedingly brilliant lustre. Hardness, 10, being the hardest known substance. Cleaves in four directions parallel to the octahedral crystal faces. Specific gravity, 3.5. Burns away completely at high temperature. Found in a peculiar kind of igneous rock in the great mines of South Africa. Also occurs in gravel deposits, as in Brazil and India.

Dolomite. Composition: calcium, magnesium, carbon, and oxygen. $CaMg(CO_3)_2$. Crystals are usually six-sided rhombohedral forms. Three

cleavage directions, much as in calcite. Hardness, nearly 4. Specific gravity 2.8. Colorless or white when pure, but variously colored by impurities. Translucent to opaque. Often difficult to distinguish off-hand from calcite, but it does not bubble when touched with a drop of cold dilute hydrochloric acid. It is a common and widespread mineral, especially in vein deposits, and in certain kinds of limestone and marble.

FIG. 394. A group of feldspar (microcline) crystals.

Feldspars. Here are included a number of mineral species and varieties, all very closely related in composition and properties. They are all silicates of aluminum with either potassium, sodium, or calcium. All have common properties as follows: Color, usually white, gray, or pink; transparent to opaque; crystals in form of prisms with faces meeting at or near 90° or 120° (Fig. 394); two good cleavages at or near 90°; hardness at or near 6; and specific gravity of about 2.6, which is the average for all minerals. The most common potash feldspar is *orthoclase* ($KAlSi_3O_8$), with two cleavages at exactly 90°. The soda-lime feldspars go by the general name of *plagioclase*. They are slightly different from orthoclase in crystal form and have cleavages at approximately 86°. Very commonly cleavage faces in one of the two directions show fine parallel lines caused by a peculiar development (called "twinning") during the crystal growth. Among the several varieties of plagioclase are *albite*, including most *moonstone*, and *labradorite*, which is usually gray to greenish gray with a beautiful play of colors. Feldspars are by far the most abundant minerals of the earth's crust. They occur in all of the three great groups of rocks—igneous, sedimentary, and metamorphic—but their most common home is in the igneous rocks, as for example granite. Potash feldspar is used in the manufacture of porcelain and chinaware. Special varieties are used as semiprecious stones or for decorative purposes.

Fluorite. Often called "fluor spar." Composition: a compound of calcium and fluorine. CaF_2. Commonly found in the form of cubic crystals. Colorless when pure, but variously, and often beautifully, colored,

especially blue, green, purple and yellow, due to impurities in solution during crystallization. Transparent to translucent. Four well-developed cleavages meeting at such angles as to permit octahedrons of very regular shape to be broken out of crystals. Hardness, 4. Specific gravity, 3.2. Very common in vein deposits, often associated with ores. Some veins in mines of southern Illinois are 20 to 40 feet wide. Mostly used in the manufacture of glass, enamel ware, and a certain kind of steel.

FIG. 395. Galena crystals.

Galena. Composition: a combination of lead and sulphur. PbS. Crystals are most commonly cubes (Fig. 395) and octahedrons. Color, lead-gray with metallic lustre which tarnishes dull. Opaque. Hardness, 2.5. Specific gravity, 7.5 (notably high). Very brittle. Three excellent cleavages at right-angles, and parallel to the cubic crystal faces. It is the one great ore of lead, being mined in many parts of the world, as for example Missouri, Colorado, Idaho, and the Rhine district.

Garnets. The term "garnet" includes a number of mineral species or varieties very closely related in composition, crystal form, and physical properties. Composition: silicates of aluminum with either iron, calcium, or magnesium. $Fe_3Al_2(SiO_4)_3$ is common. Crystals are thickset, usually with 12 or 24 faces, or a combination of the two (Fig. 396). Cleavage, scarcely noticeable. Hardness, 6.5 to 7.5 (extra high). Specific gravity, 3.1 to 4.3. Hardness and specific gravity vary according to species. Color varies with composition; but mostly red, brown, black, or green. Transparent to opaque. Garnets are mostly found in metamorphic and igneous rocks. Used as a precious stone, and also in the manufacture of so-called "garnet paper" which is similar to "sandpaper."

Gold. Known as "native gold." Composition: pure gold. Crystals are usually thickset, octahedral forms, but they are rare. No cleavage. Color, characteristic gold-yellow. Opaque. Hardness, less than 3. Specific gravity, over 19, being exceedingly high. Extremely malleable. Gold is, in small amounts, very widely distributed. Most of it occurs in gravel ("placer") deposits, and in vein deposits.

Graphite. Commonly called "black lead," but it is not lead at all. Composition: pure carbon—the same as that of the diamond, but with strikingly different physical properties. Color, black with metallic lustre. Opaque. Crystallizes in thin, flexible, six-sided plates or flakes. Hardness, between 1 and 2. Easily rubs off on paper, and feels greasy. Specific gravity, 2.2. The most natural home of graphite is metamorphosed (crystallized) sedimentary rocks. Also occurs in veins, and in some igneous rocks. It has many uses, among them being as a lubricant, in making "lead" pencils, crucibles, graphite paint, stove polish, etc. Mined in northern New York, Pennsylvania, Ceylon, etc.

FIG. 396. Garnet crystals in schist.

Gypsum. Composition: a compound of calcium, sulphur, and oxygen containing water in combination. $CaSO_4.2H_2O$. Crystals, usually tabular prismatic (Fig. 8m). Colorless or white when pure. Transparent to opaque. Hardness, 2, and easily scratched by the finger-nail. Specific gravity, 2.3. Three good cleavages crossing at angles of 66° and 114°. Moderately flexible in thin plates. A very clear, crystalline variety is called *selenite;* a fibrous variety, *satin spar;* and a massive or granular variety, *rock gypsum.* Gypsum is a common and widespread mineral, especially associated with sedimentary rocks in the form of layers and veins. Its greatest uses are in the manufacture of Plaster of Paris and (when burned) as a retarder for cement.

Halite. Usually known as "common salt." Composition: a compound of sodium and chlorine. NaCl. Crystals are usually cubes. Hardness, 2.5. Specific gravity, 2.5. Colorless to white when pure. Translucent. Characteristic salty taste. Three good cleavages at right angles and parallel to faces of the cubic crystal. Abundant and widespread in sedimentary formations of nearly all ages, sometimes as beds of *rock salt* and sometimes as natural brine or veins. Vast quantities occur in the sea and in salt lakes. Halite is very useful, as for example for cooking and preservative purposes, indirectly in glass and soap making, for glazing pottery, and in many ore-smelting and chemical processes.

Hematite. Composition: a combination of iron and oxygen. Fe_2O_3. Crystallizes in rather complex six-sided forms (Fig. 8n). Often in rounded masses (Fig. 397). Color when crystalline is black with metallic lustre, otherwise it is dull red. Opaque. Streak is always red. No cleavage. Hardness, about 6. Specific gravity, about 5 (notably above the average). It is very widespread in rocks of all ages, especially in metamorphic and sedimentary rocks, in both beds or layers, and in veins. It is the greatest iron ore, being extensively mined in Minnesota (Fig. 384), Michigan, Wisconsin, and Alabama.

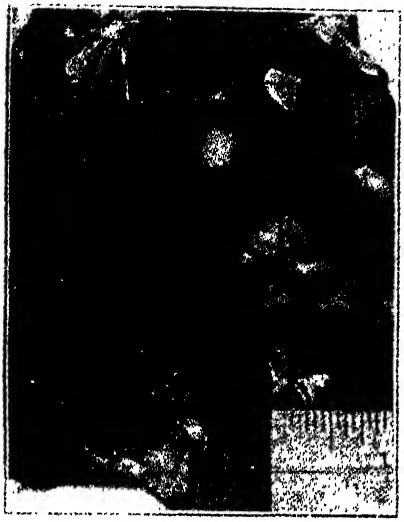

FIG. 397. Rounded masses of hematite.

Kaolin. Commonly called "china clay." Composition: a silicate of aluminum and hydrogen. $H_4Al_2Si_2O_9$. Seldom crystallizes in a small scalelike form. Usually occurs in compact, claylike masses. Color, white when pure. Translucent to opaque. Usually feels smooth and plastic. Hardness, over 2 when crystallized; otherwise it is softer. Specific gravity, 2.6. Kaolin forms the main body of much clay and shale, and so it is very widespread and abundant. Usually results from decomposition of feldspar. Pure deposits are worked for manufacture of chinaware, pottery, porcelain, etc.

Limonite. Composition: a compound of iron and oxygen similar to hematite, but also it contains water. Usual composition, $2Fe_2O_3.3H_2O$. Never crystallizes. Color, black to light and dark brown. Opaque. Gives a characteristic yellowish brown streak. Hardness, about 5. Specific gravity, nearly 4. Very common and widespread. Always a product of decomposition of various iron-bearing minerals. It is an iron ore of some importance.

Magnetite. Composition: a compound of iron and oxygen in different proportions than in hematite. Fe_3O_4. Crystals, usually in regular octahedral forms. Color, black. Opaque. Streak, black. Highly magnetic. Hardness, 6. Specific gravity, 5. Widespread as crystals in all kinds of igneous rocks, and in some metamorphic rocks. Also occurs as more or less irregular large masses and layers in certain igneous and metamorphic rocks, and in some sands. It is an important iron ore, as for example in northern New York, Norway, and Sweden.

Malachite. This mineral is in almost every way like azurite, except for its color (green), and a slight difference in composition. Translucent to opaque. It is an important ore of copper, as for example in Arizona, New Mexico, Chile, and Mexico.

Micas. Several species closely related in composition and properties are here included. Composition: complicated silicates of aluminum usually with either potash, hydrogen, magnesium, or iron. Crystals are six-sided plates or prisms whose angles are almost 120°. One very fine cleavage at right angles to the prismatic faces, yielding exceedingly thin, elastic sheets. Hardness, 2 to 2.5. Specific gravity, 2.7 to 3. Transparent to opaque. Hardness, weight, and color vary with the species, most common of which are *muscovite* or *isinglass,* which is colorless in thin sheets where pure; *biotite,* which is black; and *phlogopite,* which is brown. Among the uses of muscovite are as insulating material in electrical apparatus, for stove fronts, as a lubricant, etc.

Olivine. Often called "chrysolite." Composition: a silicate of iron and magnesium. $(MgFe)_2SiO_4$. Crystals are usually stout prismatic forms. Color, usually yellowish green. Transparent to translucent. Hardness, nearly 7 (extra high). Specific gravity, 3.3. No good cleavage. It occurs most commonly in dark, iron-bearing igneous rocks. A clear, green variety, called *peridot,* is used as a gem stone.

Opal. Composition: a compound of silicon and oxygen containing water in varying amount. $SiO_2.nH_2O$. Never crystallized, probably because of its indefinite composition. Hardness, 5.5 to 6.5, varying with varieties. Specific gravity, about 2. Color, variable. Transparent to opaque. A few of its varieties are *common opal,* and *wood opal,* usually white to light brown, translucent, and with a greasy lustre; *precious opal,* translucent with a beautiful play of colors and used as a gem; and *hyalite* in colorless, rounded masses.

Platinum. Known as "native platinum." Thickset crystals are very rare. Composition: pure platinum. Color, light steel-gray with metallic lustre. Opaque. Hardness, 4.5 (high for a metal). Specific gravity when pure, over 21, making it one of the very few of the heaviest known substances. Very malleable. Found mainly in gravel ("placer") deposits, and rarely in certain igneous rocks, mostly in the Ural Mountains of Russia. Used in the electrical industry, for jewelry, and in making certain scientific instruments.

Pyrite. Commonly called "iron pyrites." Sometimes called "fools' gold." Composition: a combination of iron and sulphur. FeS_2. Crystals are usually cubes, or thickset twelve-faced forms (Fig. 8p). Color, light brass-yellow with metallic lustre. Streak, greenish black. Opaque. Practically no cleavage. Hardness, 6, or greater than that of steel. Specific gravity, about 5 (notably higher than the average). Lighter colored and

much harder than chalcopyrite. Common and widespread in rocks of nearly all kinds and ages, but it especially occurs as veins and lenslike deposits in certain metamorphic rocks, as in Virginia, northern New York, and Spain, where it is mined. Used mostly in the manufacture of sulphuric acid ("oil of vitriol").

Pyroxenes. A number of species closely related in composition and properties are here included. In most respects, particularly in composition, the pyroxenes are much like the amphiboles (above described). Crystals are usually stout to thick-tabular prismatic forms (Fig. 8q), with the principal faces making angles of approximately 87° and 93° instead of approximately 56° and 124° as in the common amphiboles. Two cleavages (not always readily seen) parallel to the prismatic faces cross at angles of 87° and 93°. Hardness, 5 to 6, and specific gravity, 3.2 to 3.6, varying with the species. Color, commonly from white through brown to black and sometimes green. Transparent to opaque. The most common species or variety is *augite,* which is dark green, or brown to black. It is a very common constituent of igneous and metamorphic rocks. Clear green *diopside* is sometimes used as a gem stone. One kind of *jade* is a pyroxene.

Quartz. Composition: a compound of silicon and oxygen. SiO_2. Unlike opal, it contains no water. Crystals are very commonly six-sided regular prisms capped by rather steep six-sided or three-sided pyramids (Figs. 6 and 8s,t). Hardness, 7, being much harder than the average mineral. Specific gravity, 2.6 (the average for all minerals). Practically no cleavage, and breaks like glass. Colorless when pure, but varieties exhibit many colors. Transparent to opaque. Among the distinctly crystalline varieties are: *rock crystal,* pure and colorless; *amethyst,* purple; *rose quartz,* pink; and *smoky quartz,* dark. Among the very fine grained and non-crystalline varieties are: *Chalcedony,* bluish gray, waxy looking, usually in rounded masses; *carnelian,* red; *prase,* green; *agate,* banded in colors; *flint,* dark and somewhat translucent; and *jasper,* red or brown and opaque.

Next to feldspar, quartz is the most common mineral of the earth's crust. At and near the surface it is the most abundant and widespread. It is common in all three great groups of rocks—igneous, sedimentary, and metamorphic. It is a very common vein-filling mineral, often associated with ores. Some of its varieties are used for ornamental and semi-precious stone purposes. It is used in making glass, sandpaper, porcelain, mortar, concrete, and in some ore-smelting processes. Sandstone, usually consisting mostly of quartz, is widely used as a building stone.

Serpentine. Composition: a silicate of magnesium. $H_4Mg_3Si_2O_9$. Color, usually light gray, yellowish green, olive-green, or blackish green, with waxy lustre. Translucent to opaque. Does not crystallize as such. Hardness, variable from 2.5 to 5. Specific gravity, about 2.6. The most common kind of *asbestos* is a fibrous, light-green to white variety of serpentine. Serpentine is always of secondary origin, being a decomposition product of

other minerals such as olivine, amphibole, pyroxene, etc. It is common and widely distributed, especially in, and associated with, certain igneous and metamorphic rocks. In large masses it is quarried as a building and decorative stone. Asbestos is much used in making various fire proof materials.

Siderite. Composition: a compound of iron, carbon, and oxygen. $FeCO_3$. In its crystal form, cleavages, hardness, and effect of warm hydrochloric acid, it is very much like dolomite. Color, light to dark brown. Translucent to opaque. Specific gravity, about 4. Usually found in layers in sedimentary formations, or in veins. Used as an iron ore, especially in Great Britain.

Silver. Known as "native silver." Composition: pure silver. Seldom well crystallized, but usually occurs as irregular masses, plates, and wire-like forms. Color, silver-white with metallic lustre. Tarnishes to dark on exposure to air. Opaque. Hardness, nearly 3. Specific gravity, 10.5 (extra heavy). Malleable. Usually occurs in vein deposits along with other metal-bearing minerals.

Sphalerite. Composition: a simple compound of zinc and sulphur. ZnS. Crystals are usually modified tetrahedrons. Color, yellowish brown, brown, to black, with resinous lustre. Transparent to translucent. Hardness, nearly 4. Specific gravity, 4. Very good cleavages in six directions crossing at angles of 90° and 120°, so that regular twelve-sided cleavage pieces may be broken out of crystals. Sphalerite is fairly common and widespread, nearly always occurring in vein deposits. Usually associated with other metal-bearing minerals, particularly galena. It is the greatest ore of zinc.

Sulphur. Known as "native sulphur." Crystals are usually combination pyramidal forms with top and bottom truncated (Fig. 8u). Color, characteristic sulphur-yellow with resinous lustre. Transparent to translucent. Hardness, about 2. Specific gravity, about 2 (unusually low). Cleavages, very poor. Extensive deposits, as in Sicily, have resulted from decomposition of certain sulphur-bearing formations, especially gypsum beds. Some is of volcanic origin. Great quantities are used in making sulphuric acid, matches, gunpowder, fireworks, and in vulcanizing and rubber goods bleaching.

Talc. Often called *steatite*. Composition: a compound of magnesium, silicon, oxygen, and hydrogen, much like that of serpentine. Tabular and flakelike crystals are rare. Color white, light gray to greenish. Translucent. Cleavage, excellent in one direction, yielding flexible flakes. Feels greasy. Can be sliced with a knife. Hardness, 1 (very soft). Specific gravity, 2.8. Always a secondary product, resulting from decomposition of certain magnesia-rich minerals. Used to weight paper, in soap, and as talcum powder. A compact, more or less impure variety, called *soapstone*, is used for making wash tubs, electrical switchboards, blackboards, stove lining, etc.

Topaz. Composition: a silicate of aluminum with fluorine. $(AlF)_2SiO_4$. Crystals are usually flattened prisms capped by pyramids at one end, and abruptly terminated at the other. Colorless when pure, but variously colored by impurities. Transparent to translucent. One cleavage only, at right angles to prismatic faces. Hardness, 8 (very hard). Specific gravity, 3.5. Usually found in cavities in igneous rocks. Topaz is a highly prized gem stone.

Tourmaline. Composition: a complex silicate mainly of aluminum and boron with varying amounts of iron, magnesia, manganese, lime, soda, potash, etc. Crystals are prisms with faces in multiples of three capped at each end by pyramids (Fig.8v). Color varies with varying composition, but mostly black or brown. Transparent to opaque. Hardness, over 7 (high). Specific gravity, about 3. Practically no cleavage. Commonly occurs in certain kinds of igneous and metamorphic rocks. Some of the transparent, colored varieties are excellent gem stones.

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